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JOURNAL OF
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VOLUME I 1937–1938 NUMBER 2

LIGHT PENETRATION IN THE CARIBBEAN SEA AND IN THE GULF OF MEXICO

By

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and

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The growing interest in the intensity of light beneath the surface of the ocean from both the physical and the biological points of view has resulted in further efforts to obtain reliable measurements under widely differing conditions and in a variety of locations. For the most part, however, these measurements have been made in northern waters and have been confined largely to coastal regions. There has resulted a paucity of precise observations in the tropical parts of the ocean. The work of Grein (1913–14) in the Mediterranean produced highly irregular results (cf. Atkins, 1932). Certain visual observations have been made by Beebe (1934) off Bermuda, and a small number of measurements were undertaken by Stephenson (1934) from a submarine off California. The work of Hulburt (1932) was confined to measurements of the absorption coefficients of samples of sea water brought into the laboratory.

To extend our knowledge of light penetration to the more southern parts of the North Atlantic, our photometric apparatus was placed aboard the research vessel "Atlantis" at the outset of her cruise to the Caribbean and the Gulf of Mexico during the winter of 1936–37. Measurements were carried out at every suitable opportunity, but because of the press of other work and the frequent occurrence of stormy weather, a much smaller number of series was completed than was hoped. The observations presented

* Contribution no. 168.

below are therefore not to be regarded in any sense as a complete survey, but represent a first step in filling a serious gap in our knowledge of light conditions in tropical and sub-tropical waters.

The deck and sea photometers employing Westinghouse Photox Cells described by Powell and Clarke (1936) were used for this investigation. The spectral sensitivity of these cells exhibits a maximum at 5500A and drops off to 10% of the maximum at 4650A and 6250A. This relatively broad sensitivity admits of the criticism that the value obtained for the

Figure 26. Map showing the locations of stations where light penetration measurements were made. The "Series Number" of each set of observations appears beside the circle representing the station.

transparency of the water is a composite of the transmissive exponents of a large number of wave lengths, the relative intensities of which may differ with depth. However, since the logarithmic plots of measurements made in homogeneous bodies of water are remarkably straight, any such discrepancies must be of negligible proportion. It would appear that, in these cases at least, the most effective radiation at all depths centers about the same part of the spectrum.

The spectral band to which the present photometers are sensitive is essentially the same as that measured by the "green-sensitive" combination of filters and photoelectric cells employed by Oster and Clarke (1935). The spectral sensitivity of the Photox cells corresponds so closely to that of the human eye that we may refer to the light measured as the visible

part of the total solar radiation. Furthermore, since approximately the same spectral sensitivity appears to characterize the photosensitive organs of marine animals generally (Clarke, 1936a), the reduction of light indicated by our instruments as they are lowered into the sea is presumably the same as that perceived by a fish or other animal sinking a corresponding distance in the water.

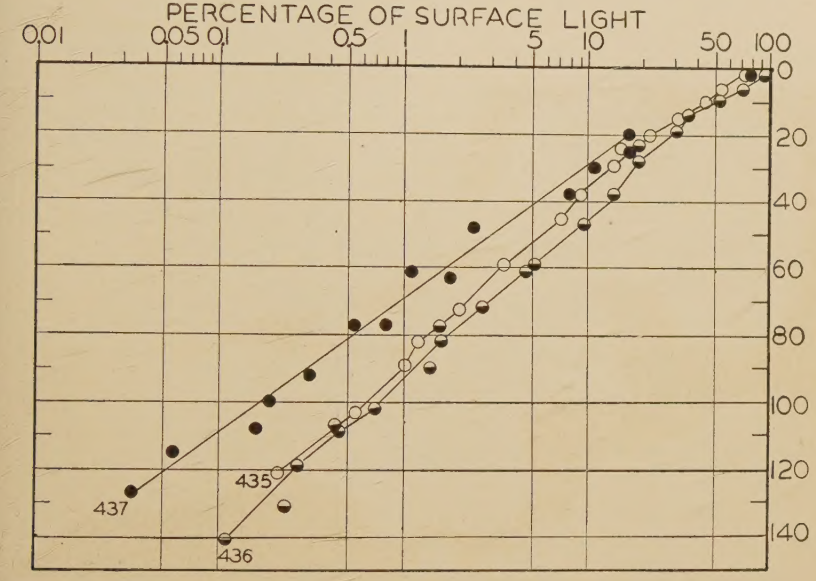

Figure 27. Relation between depth and irradiation expressed as a percentage of the light just over the surface (logarithmic scale). Series numbers indicated at end of each curve.

Series	Locality	Latitude	Longitude	Date	Time (E.S.T.)	Sky*	Sea	Wind†
435	Antilles Current	24°00′N	59°03′W	Jan. 4, 1937	09:50–10:44	bc	moderate	3–4
436	No. Equatorial Current	12°04′	60°29′	Jan. 12, 1937	10:55–11:45	o	rough	4–5
437	Caribbean Sea	15°20′	67°10′	Jan. 26, 1937	10:00–10:42	bc	moderate	4–5

* U. S. Weather Bureau symbols. † Wind force on Beaufort Scale.

The same procedure was followed and the same precautions observed in making these light penetration measurements as has been described previously. I am indebted to Mr. Alfred Woodcock for carrying out the observations at sea and to Mr. David Bonnet for assistance in preparing the figures.

The location of each series of measurements is shown in Fig. 26. Series 435 was made in the Antilles Current at a distance of about 900 miles SE of Bermuda. Series 436 was located in the Northern Equatorial Current just East of the Lesser Antilles. Series 437 and 438 were made respectively

in the center of the Caribbean Sea and in the Cayman Sea* just West of Jamaica. Series 439 to 442 were made in the Gulf of Mexico East of the Mississippi Delta at different distances from shore.

TABLE I

COMPARISON OF AVERAGE VALUES OF THE TRANSMISSIVE EXPONENT, k,* FOR THE "GREEN" COMPONENT OF DAYLIGHT FOR THE DEPTH RANGES INDICATED

Series	*Station*	*Locality*	*Ranges (meters)*	k	*Total Range (meters)*	k
307	2243	Sargasso Sea, N of Bermuda	—	—	5–92	.050
311	2245	Gulf Stream, N of Bermuda	—	—	5–136	.049
415	2484	Gulf Stream, S of Grand Banks	2–10	.082		
			10–95	.057	2–95	.059
435	2726	Antilles Current	2–20	.068		
			20–121	.045	2–121	.049
436	2748	Northern Equatorial Current	2–14	.079		
			14–141	.045	2–141	.048
437	2803	Caribbean Sea	20–130	.058†		
438	—	Cayman Sea, W of Jamaica	2–95	.045		
			95–185	.038	2–185	.042
439	—	Gulf of Mexico, E of Mississippi Delta	2–25	.22		
440	—	Gulf of Mexico, E of Mississippi Delta	2–20	.19		
			20–53	.10		
441	—	Gulf of Mexico, E of Mississippi Delta	2–6	.16		
			6–24	.088		
			24–81†	.058†		
442	—	Gulf of Mexico, E of Mississippi Delta	2–95	.054		
			95–169	.039	2–169	.047

* In the equation $I/I_0 = e^{-kL}$ where L is the thickness of the water layer (in meters) in which the intensity is reduced from I_0 to I.
† Approximate.

The transparency of the water at the station in the Antilles Current (Series 435) and at the station in the Northern Equatorial Current (Series 436) was found to be uniform over the ranges measured, as is attested by the straightness of the logarithmic plots (Fig. 27), although the upper strata tended to be somewhat less transparent than the deeper layers (Table I). The water in these areas is known to be highly homogeneous from a hydrographic point of view (Iselin, 1936; Parr, 1937), and at the time of the present observations the temperature change from the surface to a depth of 150 m. was only about 4° C. Since the average values of the transmissive exponents for these series (k = .049 and .048 respectively) are so nearly the same, we may conclude that there was practically no difference in transparency at these two localities.

* Name proposed by Parr (1937) to designate the body of water between the Gulf of Mexico and the Caribbean Sea proper (see Fig. 26).

The observations in the Caribbean Sea (Series 437) unfortunately occurred during unfavorable weather conditions with the result that the measurements are rather widely scattered. However, a straight line con-

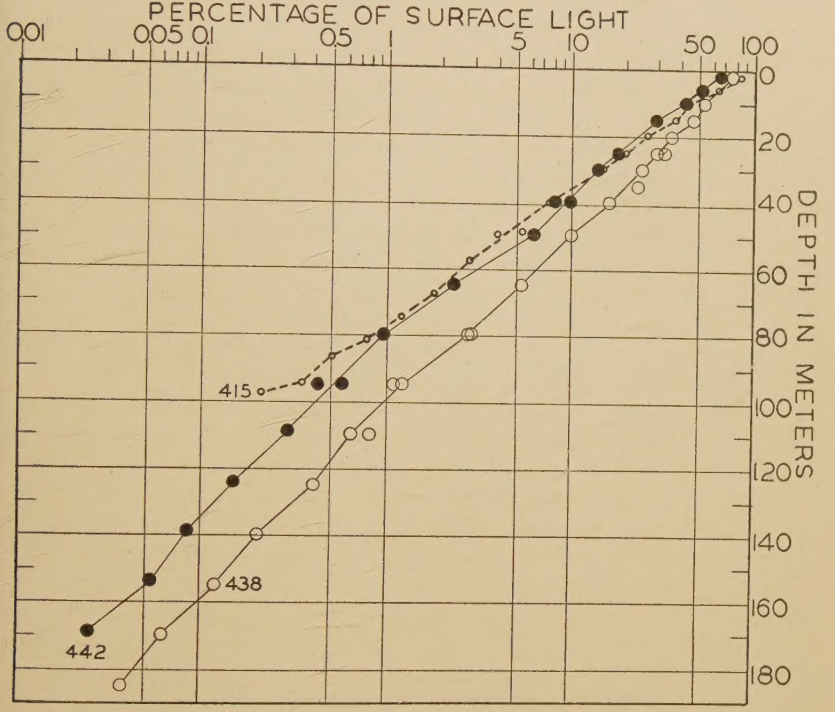

Figure 28. Relation between depth and irradiation expressed as a percentage of the light just over the surface.

Series	Locality	Latitude	Longitude	Date	Time	Sky	Sea	Wind
415	Gulf Stream (S of Grand Banks)	39°56′N	48°40′W	Sept. 15, 1935	15:35–17:00 (L.A.T.)	o	slight	3–4
438	Cayman Sea	18°38′	79°12′	Feb. 28, 1937	10:45–12:42 (E.S.T.)	bc	smooth	1
442	Gulf of Mexico	29°14′	87°48.5′	April 11, 1937	10:30–11:18 (C.S.T.)	b	smooth	1

necting the value at 20 m. with that at 130 m. appears to form a satisfactory average curve for the whole series below 20 m. The slope of this curve gives a value of $k = .058$. Although individual measurements above 20 m. were unreliable, the transparency was evidently considerably reduced in the upper stratum. The temperature at the surface was 26.0° C. and at 150 m. was 19.8° C. at this station.

Since the measurements at the station in the Cayman Sea (Series 438) were made under favorable conditions, and since the values obtained are so consistent, we may place considerable reliance upon them. The series extended to a greater depth (185 m.) than has ever been reached previously with photoelectric or rectifier cell photometers. The curve for this series (Fig. 28) shows that the water at this locality was extremely uniform in transparency as would be expected from the hydrography (Parr, 1937). From 95 m. to 185 m. the value of the transmissive exponent was found to be $k = .038$. This water is therefore the most transparent (in respect to the "green" component of daylight) ever reported (cf. Oster and Clarke, 1935) and approaches the clearness of distilled water (for which $k = .024$ at wave length 5400A according to Sawyer, 1931). If we assume that the optical properties of the water at this station are the same as those of the Sargasso water off Bermuda, which is similar hydrographically, we may conclude that the "blue" component of daylight penetrates even more effectively than the "green" in this locality and therefore, that "blue" light is of measurable intensity at an even greater depth.

The region east of the Mississippi Delta was unfortunately the only part of the Gulf of Mexico where there was opportunity to make light penetration measurements. The four series of observations here were made in increasing depths of water and show how the effect of the muddy Mississippi water diminishes at increasing distances from shore (Fig. 29). The water in Series 439, where the depth was 25 m., was quite homogeneous and gave an average transmissive exponent of $k = .22$. This water was therefore approximately as turbid as Vineyard Sound or Provincetown Harbor (cf. Clarke, 1936b). In Series 440 (depth 63 m.) we see that a distinct change in transparency occurred at 20 m. Above 20 m. the water was only slightly more transparent than at the previous station but below 20 m. the value of the transmissive exponent was reduced to one half ($k = .10$).

The measurements composing Series 441 were made under conditions causing the ship to drift more than usual. The values obtained down to 40 m. appear sufficiently consistent to be reliable but below 40 m. the points are seen to scatter widely. At these greater depths the wire supporting the submarine photometers departed by as much as 60° from the vertical, making a precise determination of the depth of each observation difficult. Incorrect estimation of depth due to wire angle is a dangerous source of error in all deep readings and should be carefully guarded against. In this case the last measurement of the series was made with the photometer actually resting on the bottom. As a result the depth could be independently and accurately determined by means of the ship's fathometer and was found to be 81 m. A straight line has been drawn on the plot (Fig. 29) from the value obtained at this depth to the value found at 38 m. This (dashed) line may serve as an approximation for the illumina-

tion below 40 m. The complete curve shows that the water in the upper six meters was still highly turbid (k = .16). A transition layer occurred from 6 to 24 m. (k = .088). Below 24 m. the water was considerably more transparent (k = .058) than at the previous station.

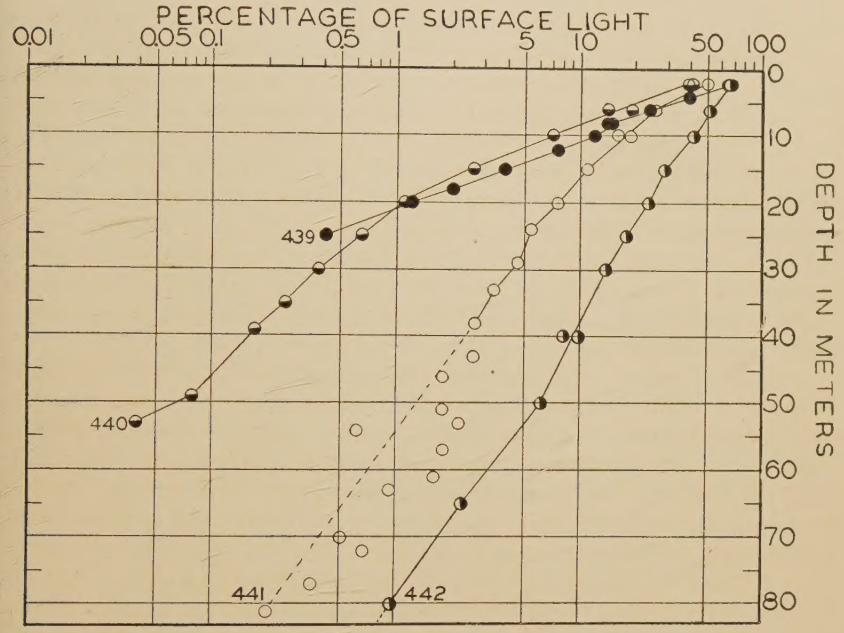

Figure 29. Relation between depth and irradiation expressed as a percentage of the light just over the surface.

Series	Locality	Latitude	Longitude	Date	Time (C.S.T.)	Sky	Sea	Wind
439	Gulf of Mexico	29°27′N	88°48′W	April 9, 1937	11:34–11:54	b	slight	3
440	Gulf of Mexico	29°16′	88°50′	April 9, 1937	15:30–15:53	bc	slight	3
441	Gulf of Mexico	29°08′	88°39′	April 10, 1937	10:45–11:30	b	slight	3
442	Gulf of Mexico	(See Fig. 3)						

Series 442 was made in 656 m. of water and under much more favorable weather conditions. A comparison with the other measurements in this region (Fig. 29) reveals a further great increase in transparency at this point. There is no evidence of relatively opaque water near the surface. However, a slight change of slope occurs at 95 m. (Fig. 28) giving transmissive exponents of k = .054 and k = .039 respectively for the curve above and below 95 m. Since the water at this station was almost as clear as that at the station in the Cayman Sea (Series 438), it appears that the effect of the muddy water from the Mississippi does not extend far in this

direction from the Delta. At this point of observation, only some seventy miles from the nearest mouth of the river, the water had a transparency practically equal to the clearest ocean water known. The explanation of this apparent paradox lies in the fact the water debouching from the Mississippi is carried almost entirely to the west (Parr, 1935, Fig. 10). In speaking of the influence of the Mississippi and the other northern drainages, Parr says (loc. cit. p. 62): "This freshening influence, however, stays confined to the bank and does not seem to affect the surface layer over the deeper waters at all, even where the bank is quite narrow as at the mouth of the Mississippi itself. It will be noticed from our map of surface salinities that the freshening influence spreads westward and has very little effect to the east of the Mississippi Delta." Our measurements of transparency support these conclusions.

For the purposes of comparison three previous measurements made in Sargasso water (Series 307, 311, and 415) have been included in Table I and the curve of Series 415 has been added to Fig. 28. These observations taken together with the deep-water series reported in the present paper provide a general picture of transparency for a large part of the western North Atlantic in which oceanic water of high salinity exists. It is seen that the transparency throughout this whole region is extremely high and quite uniform. A gradual, though slight, decrease in transparency does occur from the Cayman Sea (Series 438) to a point in the Gulf Stream, south of the Grand Banks (Series 415), as is revealed by the increasing values of the average transmissive exponents (Table I and Fig. 28). However, the ratio of the highest average value of the exponent to the lowest is only .059 (Series 415)/.042 (Series 438) = 1.4. The extreme change over the whole length of the area (over 2000 miles) is accordingly only 40% in contrast to changes of 600% (for the corresponding "green" component of daylight) found by Clarke (1936b) in passing off-shore from Woods Hole toward Bermuda. Since in the latter case measurements in the turbid coastal waters are included, the two situations are not really comparable. However, most of the investigations of light penetration made heretofore—such as those of Poole and Atkins, Pettersson, and Utterback—have given a picture of great variability. The present observations are the first in which opportunity has arisen to study the transparency of a considerable oceanic area and, in this case, show a wide and uniform distribution of extremely clear water.

SUMMARY

Measurements of light penetration using Photox rectifier cells were made at 8 stations in the Caribbean Sea region and in the Gulf of Mexico. At the stations in shallow water east of the Mississippi Delta, considerable turbidity was encountered in the surface layers. But at the offshore station in the Gulf and at all the other stations the water was found to be highly uniform and extremely transparent. The value of the transmissive exponent from 95 to 185 m. at the station in the Cayman Sea west of Jamaica was $k = .038$,—indicating the presence of the clearest ocean water ever measured.

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A BATHYTHERMOGRAPH

By

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In the study of the homogeneous layer in the ocean, Rossby[1] found it desirable to have an instrument which would provide a continuous record of temperature against pressure in the surface layers of the ocean. A preliminary instrument named an "oceanograph" was constructed and used during the summer of 1934. The manifold uses to which such an instrument could be put presaged a widespread employment of the apparatus. This, however, did not come about because of certain inherent difficulties in Rossby's design. The record was made on a large smoked foil, and thus entailed the attachment of multiplying linkages to the actuating elements for pressure and temperature. Such multiplying linkages are uncertain in action in sea water, and, furthermore, the size of the instrument to accommodate them must necessarily be fairly great. At Prof. Rossby's suggestion, the author attempted to modify the oceanograph so that it would be more suitable for routine use. The modifications were made with the following aims in view:

(*a*) The instrument should if possible be small enough so that it can be lowered on an ordinary log line by hand if necessary, thus enabling it to be utilized on vessels not equipped with a hydrographic winch.
(*b*) The instrument should be sufficiently rapid in its response such that regardless of the rate at which it passes through the water no errors due to the lag of the thermometric element will be apparent.
(*c*) Care should be taken to eliminate hysteresis of the pressure element.
(*d*) The plates on which the record traces are made should be easily inserted and removed and easily evaluated.

DESCRIPTION OF THE INSTRUMENT

The instrument finally evolved consists essentially of a pressure element comprising an hermetrically sealed sylphon inside of which is a guide and compression spring to give the requisite pressure range. Mounted on the movable end of the pressure element is a straight bi-metal strip, and thus

*Joint contribution from Woods Hole Oceanographic Institution (No. 167) and New York University.

[1] C.-G. Rossby and R. B. Montgomery. "The Layer of Frictional Influence in Wing and Ocean Currents." Papers in Phys. Oceanography and Meteorology of the Mass. Inst. of Tech. & Woods Hole Oceanographic Inst., Vol. III, No. 3, p. 73.

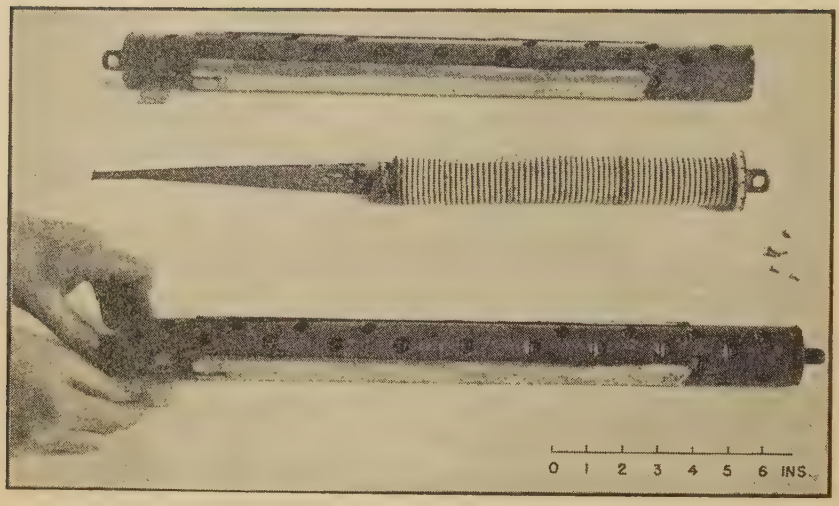

Figure 30. The bathythermograph.

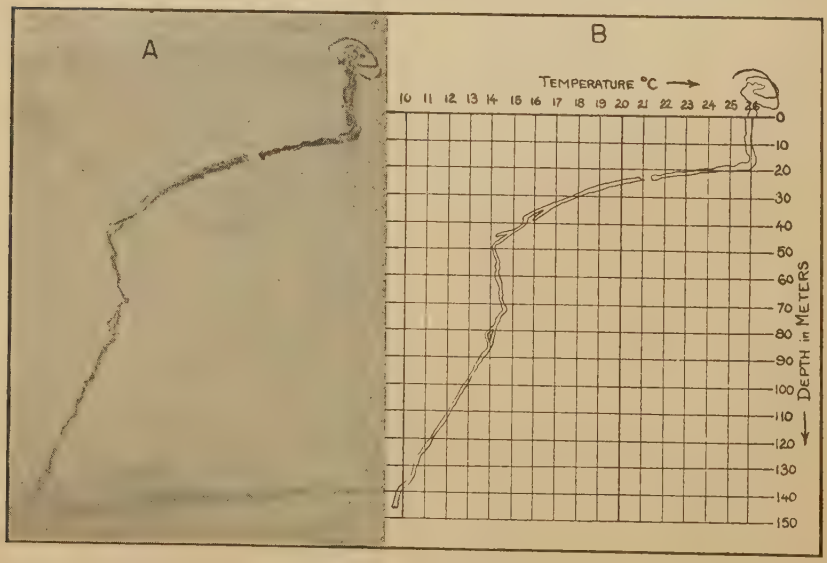

Figure 31. Photographic enlargement (A) from an original record of the bathythermograph, with key diagram (B) showing trace superimposed upon calibration chart.

motion with pressure is at right angles to the deflection of the strip with temperature. At the end of the bi-metal strip a fine needle point is arranged to inscribe a trace on a small glass slide. The slide is prepared for the record by coating it with a thin layer of oil and then smoking the glass to slightly blacken it. The function of the oil is to prevent the smoked film being washed off by the sea water. In view of the fact that the motion of the inscribing point is in two dimensions, the plate which receives the record is held perfectly rigidly in the body of the instrument, and thus insertion and extraction of the plates are entirely independent of the sensitive actuating elements. Figure 30 shows the complete instrument in the lower photograph and indicates the method of insertion of the glass slide to take the trace, while the upper part of the illustration gives a view of the instrument partly disassembled to show the simplicity of the actuating elements. Figure 31 is a photographic enlargement of an original record from the small slide, and the accompanying key diagram shows the trace superimposed on a calibration grid. It can be seen that the record consists of two lines—one made during the descent of the instrument, and the other during the ascent. It can be easily shown that hysteresis of the pressure element and lag of the temperature element both act in such a way as to separate the ascent and descent traces; so that if either pressure hysteresis or thermal lag are present to a marked degree, two distinct traces would have been recorded. The coincidence of the two curves, however, insures that the instrument is functioning correctly. From the standpoint of thermal lag the record is particularly remarkable when it is realized that in the case of the particular test in the photograph the instrument was sent down and brought up through the water at a rate of 100 meters per minute. Thus the whole test down to a depth of 150 meters was completed in 3 minutes. Evaluation of the record obtained is simply accomplished by projecting it by means of a miniature camera projector on to a calibrated screen. Tests were made by lowering the bathythermograph on the end of a log line and it was found that good records could be obtained by an ordinary seaman.

SOME RESULTS OF MEASUREMENTS MADE WITH THE BATHYTHERMOGRAPH

On a cruise made by the "Atlantis," research ship of the Woods Hole Oceanographic Institution, measurements with the bathythermograph were carried out almost hourly for a period of two days, and remarkably sudden discontinuities in the thermocline and inversions of the thermocline were revealed. Figure 32 is a plot showing the temperature variation with depth on the horizontal and vertical axes respectively, while the time changes are set off along the oblique third axis. The bathythermograph records were traced directly on to this diagram from photographic enlargements of the original traces. This figure served to illustrate several points.

(*a*) Abrupt changes in the thermocline are shown, and it should be stressed that these changes must be actually present because they are in every case substantiated by the coincidence of the up and down trace of the bathythermograph.

(*b*) The temperature depth structure below 50 meters is quite complicated, especially for those tests between 1900 and midnight. Thus it is

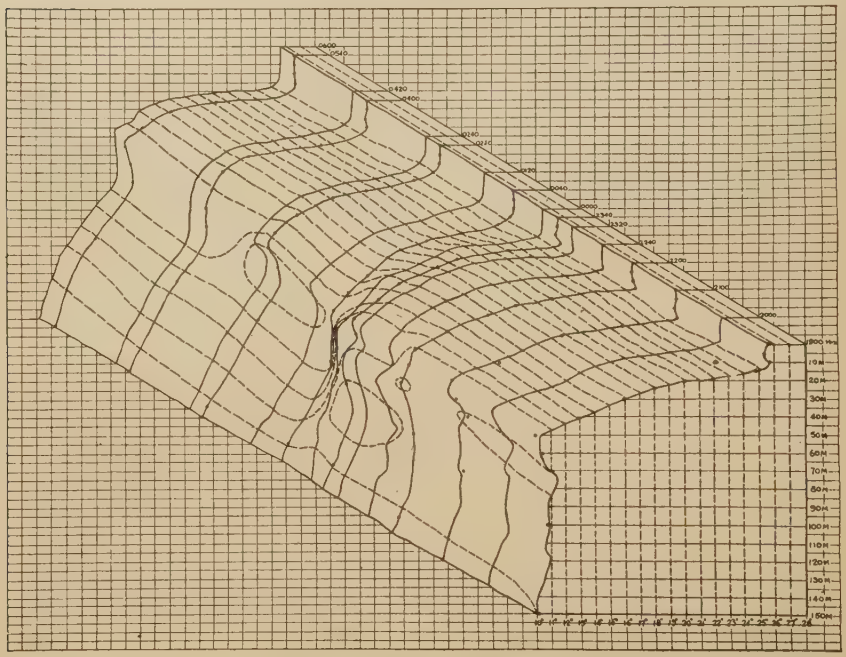

Figure 32. The variation of the temperature-depth profile with time, as determined from the bathythermograph traces and water bottle temperatures.

evident that if water bottles were used these intricacies of the temperature curve would be entirely overlooked, owing to the necessarily wide spacing of the bottles below 50 meters.[2]

[2] This was actually the case. At Station no. 18 (Aug. 26, 1937, 2203 hrs.) of the cruise on which the instrument was tested the water bottles were spaced at 1, 5, 10, 15, 20, 40, 70, 150, 200, 300, and 400 meters and the smooth curve, drawn by Mr. C. O'D. Iselin through the points thus obtained, completely missed an abrupt change from the summer thermocline to isothermalcy which the bathythermograph indicated to exist at about 60 meters. Accordingly, at Station no. 19, one hour and forty minutes later in the same water, the spacing was altered to: 1, 5, 10, 15, 20, 25, 30, 40, 50, 60, 70, 100, 150, 250, and 350 meters and the discontinuity of the thermal

(*c*) The sudden change in the character of the temperature depth curve below 50 meters between the hours 0000 and 0040 is very remarkable, for the "Atlantis" could not have drifted very far in this short space of time.[3]

(*d*) The traces shown at the following hours were taken with the bathythermograph. 1900, 2000, 2100, 2240, 2320, 0000, 0120, 0240, 0420, and 0600, while the curve obtained from water bottle thermometer readings are shown at 1900, 2100, 2200, 2340, 0040, 0220, 0400, and 0540. Thus, comparison between the bathythermograph readings and the thermometer readings may be made directly for the two times 1900 and 2100 when both series were taken nearly simultaneously.

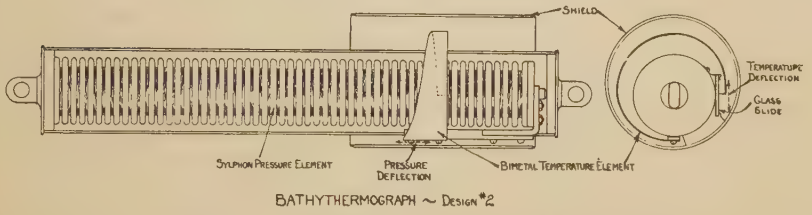

Figure 33. Design No. 2 for improved bathythermograph.

Such small differences (never more than 0.2° Centigrade) as are evident may easily be accounted for by the difference in time between the water bottle temperature station and the bathythermograph test.

FURTHER DEVELOPMENT:

Though the result shown may be considered as very satisfactory, it is proposed to improve the instrument further by an attempt to utilize a design such as indicated in Figure 33. The improvement which such a design would incur is that the temperature element would be exposed directly to the flow of water and that all thermal lag by it would be eliminated. Furthermore, the size of the instrument, if constructed according to the design in Figure 33 will be cut down considerably and, though in the first design the coordinates are very nearly rectangular (because the length of the bi-metal strip is great compared to the deflection of its end), in the

gradient as indicated by the bathythermograph was established. The latter distribution of water bottles was thereafter maintained for subsequent stations in the same region.

[3] In a personal communication, Mr. Iselin informed the author that the T-S correlation also changes at the same time and, as the drift of the ship could not have been more than a mile or two, he concludes that the mixing in the neighborhood of 100 meters depth was very incomplete.

second design the coordinates will be perfectly rectangular. Finally it is thought that vibration will be entirely eliminated by the second design, and that, therefore, a finer trace and greater accuracy will be obtained.

It is hoped that this instrument can be produced, if demand is sufficient, cheaply enough to obtain it widespread application, not only by biologists and oceanographers, but also in the fishing industry—the apparatus being such that it can be handled by entirely untrained personnel.

ON THE OXIDATION OF ORGANIC MATTER IN MARINE SEDIMENTS BY BACTERIA

By

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New Jersey Agricultural Experiment Station

and

New York Medical College and Flower Hospital

The question as to whether the organic matter in the sea bottom is inert or is available as a nutrient for the bottom fauna and bottom-inhabiting bacteria is a problem of special interest in the cycle of life in the sea. A number of marine animals, making up the benthos or the bottom fauna, must depend upon this organic matter for their nutrition while the sea bottom also harbors an extensive bacterial population which require the energy to be derived from various nutrient elements for cell synthesis and respiration.

It has been definitely established that different marine sediments vary considerably in content of organic matter, depending on such factors as the nature of the bottom material, its distance from land, the topography of bottom, and the depth of water (*11*). Little is known as to whether this organic matter can be further decomposed once it has settled to the bottom. In his studies on the decomposition of Zostera plants in the sea, Boysen-Jensen (*4*) found that there was first a marked reduction in the nitrogen content to be followed later by a gradual increase in nitrogen. This led him to conclude that non-nitrogenous substances are decomposed to a greater extent than the nitrogenous matter by the bacteria in the sea bottom. The bottom fauna itself seemed to digest largely the non-nitrogenous bodies, such as the pentosans, while the nitrogenous compounds accumulated. The nitrogenous substances were considered to be transformed to humic compounds, which are attacked much less readily; and so the rôle of the bacteria in building up the complex organic nitrogenous compounds out of the mineralized forms of nitrogen was assumed to merit consideration. Boysen-Jensen separated the bottom sample into the "brown layer," or the uppermost 1 cm., and the lower 2 cm. layer; by digesting the sediment with pancreatic enzymes, he found that the upper layer of 120 sq. cm. surface of sea bottom contained 44–68 mg. of digestible

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nitrogen and the lower layer usually none. This method of determining how available the nitrogenous constituents of the sea bottom may be is open to criticism, because, under natural conditions, the digestion process is carried out by bacteria or by benthotic animals which will digest the carbohydrates even more readily and the small amounts of nitrogen consumed by these organisms would be built up into their cell bodies rather than be liberated in a mineralized form; a fact which was recognized by Boysen-Jensen.

From the bacteriological point of view, it is important first of all to determine to what extent bacteria are able to attack the humus complexes in the bottom material and how rapidly the essential nutritive elements, especially the nitrogen, are made again available for plant synthesis. The ratio of carbon to nitrogen in the marine humus has been found (*4*, *11*) to be more or less constant, varying from 8:1 to 12:1. Under purely artificial conditions of decomposition, for instance when marine mud was air dried, remoistened and placed in a glass vessel, it was found (*11*) that 150 gm. of mud gave off, in 14 days, 55.8 mg. of CO_2 as carbon and 11.1 mg. of ammonia as nitrogen; this showed that the organic matter itself was not absolutely inert, but could be gradually, even if only very slowly, decomposed.

The above experiments carried out by the use of different methods, although artificial in nature, pointed definitely to the fact that the organic matter in the marine bottom is not absolutely inert but can be slowly made available. It remains to be determined to what extent this organic matter can undergo decomposition in a sea water system, by the natural bacterial population of the sea, and how this process affects the important elements carbon and nitrogen in the organic matter of the sea bottom.

Among the different methods used in studying the decomposition of small amounts of organic matter in sea water, the consumption of oxygen has been found to be most convenient (*12*). Therefore, it might be possible to utilize the rate of oxidation of the organic matter in the sea bottom as a measure of its availability. The application of this method to the study of certain types of bottom material, notably muds, is complicated by the fact that these can also absorb oxygen chemically. Moore, for example, observed (*9*) that there was an increase in the amount of immediate chemical absorption of oxygen by marine mud, to a depth of 5 cm.; but below that depth the chemical absorption of the oxygen became constant and amounted to 0.3–0.5 mg. of oxygen per 1 gm. of dry mud. He further stated that the biological oxygen absorption was slower, although no data were given to substantiate this. ZoBell and Anderson (*15*) reported that 1 gm. of dry mud absorbed from sea water 0.2–4.8 mg. oxygen in 6 hours, but biological absorption of oxygen continued at a slower rate almost indefinitely; both chemical and biological oxygen absorption were found to be much greater in the surface layers of mud than in the deeper layers. No details were

given concerning the method for determining the biological absorption of oxygen, while in both contributions even the chemical absorption was only vaguely defined and described.

Various studies are on record concerning the absorption of oxygen by bottom material in lakes and in other bodies of water. Alsterberg (*2*) suggested that the disappearance of oxygen from the depth of a lake is due to the presence of easily oxidizable organic substances, often planktogenic in nature, in the bottom material. Kusnetzow (*7*) explained the disappearance of oxygen from the deeper portion of a lake as a result of the activities of the bacteria oxidizing methane, which had its origin in the anaerobic decomposition of the organic matter in the bottom material. He further reported (*8*) that the oxygen changes in a lake could be used as a measure of the intensity of the microbiological processes taking place in the bottom material.

Methods. The measurement of oxygen consumption has been of value in previous biological studies from this laboratory and it was decided that this method should be adapted to the present problem. In brief, this method consisted in the addition of the material to be studied to identical glass stoppered bottles which are incubated, usually submerged, at a predetermined temperature. At definite incubation times duplicate bottles were removed, the oxygen determined by the Winkler method, and the results expressed in cubic centimeters of oxygen per bottle. The bottles used in this work contained 225 cc. of liquid.

In attempting to use this method for the oxidation of marine bottom material, the method had to be variously modified: (1) It was necessary to compare the specific behavior of different types of bottom material. (2) It was desirable that the water in which the mud was suspended should be relatively inert and not exert too great an influence on the oxygen balance. (3) It was essential to determine how great an effect on oxygen consumption was exerted by the non-biological constituents of the bottom material. (4) The Winkler method being indirect, it was of prime importance to make sure that the results obtained were not due to chemical interactions of the reagents and the samples used.

The bottom material was obtained in a sterile glass tube which was lowered in a brass tube surrounded by a heavy weight; this is similar to the corer employed by Moore and Neill (*10*). The glass tubes with the mud cores were stoppered at each end with sterile rubber stoppers, kept on ice on the ship, and used as soon as possible after their return to the laboratory. The cores were carefully separated into two layers; the surface, brownish, freshly sedimented layer designated as the A horizon, and the sub-surface darker layer or B horizon. Quantities of mud varying from 0.1 to 0.5 gm., on a dry basis, were used; while for the sand 2 to 5 gm. portions were added to the 225 cc. bottles. The dry content of the material was calculated from

a separate aliquot sample taken for moisture determination. The final results were recorded on the basis of one gram portions of dry material.

Although it was possible to use sea water newly procured as a menstruum in which to suspend the mud, the rapid oxygen consumption resulting from the oxidation of the organic matter in such sea water (*12*) might have been confused with the oxygen consumption due to the added bottom material. The sea water menstruum was therefore stabilized by ageing. This was accomplished by securing a quantity of water, taken from Woods Hole Harbor at a point distant from the pier, filtering it through a fine silk net (#20) and placing it in 18 liter glass bottles in the laboratory for at least 5 days. Before use the water was aerated to re-saturate with oxygen.

A series of experiments with sterilized and unsterilized material showed that there was very little oxygen consumption under aseptic conditions and gave ground for the belief that "chemical" oxidation was of little significance as compared with "biological." As a consequence, the containers with mud and water were placed under conditions favorable to bacterial activity. Room temperature was used, and oxygen determinations were made at the start and throughout the duration of the experiment, which was usually 5 to 15 days. When bacterial numbers were determined, by the plate method, a sample of water was usually taken above the settled bottom material.

After a few determinations had been made by adding the Winkler reagents directly to the bottles of water and mud or water and sand, it was apparent that the oxygen values obtained reflected an oxygen consumption in excess of the probable biological consumption. This was because the mud particles reacted with the iodine liberated. To eliminate this interference, the supernatant liquid was obtained by siphoning into smaller oxygen bottles (130 cc. capacity) and these were used for the oxygen determinations. It was found that siphoning did not interfere with the oxygen content of the water. The supernatant liquid from above the mud always gave a higher oxygen content than similar material which had not been freed of mud. The difference was sometimes as little as 0.03 cc. oxygen or as great as 0.19 cc. oxygen per bottle.

Experimental. In the course of the development of a standard method to be used in handling fresh bottom materials, data concerning several factors of interest in laboratory experiments on decomposition were gradually collected.

One of these preliminary experiments showed the extent of oxygen consumption which might be expected with dried material. Two dried mud samples, which had been previously obtained (*11*) from *Atlantis* stations 1329 and 1331 and found to contain 2.45 and 1.46 per cent organic carbon and 0.285 and 0.143 per cent nitrogen respectively, were studied. The oxygen consumption during 54 hours was compared with that of dry sand

bottom from George's Bank containing 0.4 per cent carbon and 0.045 per cent nitrogen. The organic matter of both mud and sand was oxidized, although not very rapidly (Table I). In the case of sample 1329, 0.5 gm. of dry mud consumed 0.14 cc. oxygen, while 2 gm. of mud consumed 0.32 cc. oxygen in 54 hours. The corresponding amounts of oxygen consumed by mud 1331 were 0.12 and 0.32 cc. Although there was a difference in the

TABLE I

OXIDATION OF DRY MARINE MUD AND SAND IN SEA WATER

Oxygen content of water, cc. per bottle

Mud Material	*Amount used*[1]	*Incubation, hours*			
	gm.	*0*	*6*	*30*	*54*
Water alone	—	1.21	1.19	1.16	1.15
1329 mud	0.5	1.20	1.18	1.10	1.01
1329 mud	2.0	1.14	1.13	0.99	0.83
1331 mud	0.5	1.18	1.16	1.09	1.03
1331 mud	2.0	1.12	1.12	0.96	0.83
Sand	2.0	1.20	1.20	1.08	0.99
Sand	5.0	1.24[2]	1.21	0.97	0.84

[1] Per 225 cc. oxygen bottle.

[2] Higher oxygen content above the control is due to a few small bubbles of air remaining in the sand.

TABLE II

INFLUENCE OF STERILIZATION UPON OXYGEN CONSUMPTION BY MARINE BOTTOM MATERIAL

Material[1]	*Oxygen, cc. per bottle*				*Bacterial control after 120 hours*
	Start 1 hr.	*18 hrs.*	*42 hrs.*	*120 hrs.*	
Standard sea water	1.07	1.08	1.07	—	—
+ 2 gm. mud 1331	0.99	0.96	0.92	—	—
Standard sea water	0.94	0.94	0.94	0.94	0
+ 3.75 gm. Gay Head sand	0.91	0.92	0.91	0.91	0
+ 1.47 sandy mud	0.94	0.92	0.91	0.88	0
+ 0.47 gm. mud	0.87	0.79	0.76	0.65	+

[1] Amounts calculated on dry basis.

organic matter content of the two muds, as shown by the carbon contents, there was very little difference in the rate of oxygen consumption. In the case of sand, however, much less oxygen was consumed: 2 gm. of sand absorbed, in 54 hours, 0.16 cc. oxygen; 5 gm. absorbed 0.31 cc.

While experiments under aseptic conditions showed that sterilized material absorbed only a very small amount of oxygen from the water (Table II), the biological oxygen absorption by the same samples presented a different picture. This is shown graphically in Figure 34. The sand used was from 22 meters depth, 1.5 miles W. by N. from Gay Head (Marthas

Vineyard); it had 24.9 per cent moisture. The sandy mud was obtained from a depth of 37 meters, near Buoy No. 4 LB, 7 miles S.W. of Gay Head; it had 26.4 per cent moisture. The mud came from Buzzards Bay, 1 mile N.E. of Gong 6; it was at a depth of 16 feet and had a moisture content of 71.0 per cent. Two gram portions of the moist sand and 0.5 gm. portions of the moist sandy mud and mud were used per 225 cc. oxygen bottles. Calculations were made on the basis of one gram of dry material. In thirty one days, 1 gm. of the sandy bottom containing 0.8% organic matter (calculated from the nitrogen content) absorbed 0.44 cc. oxygen; the sandy

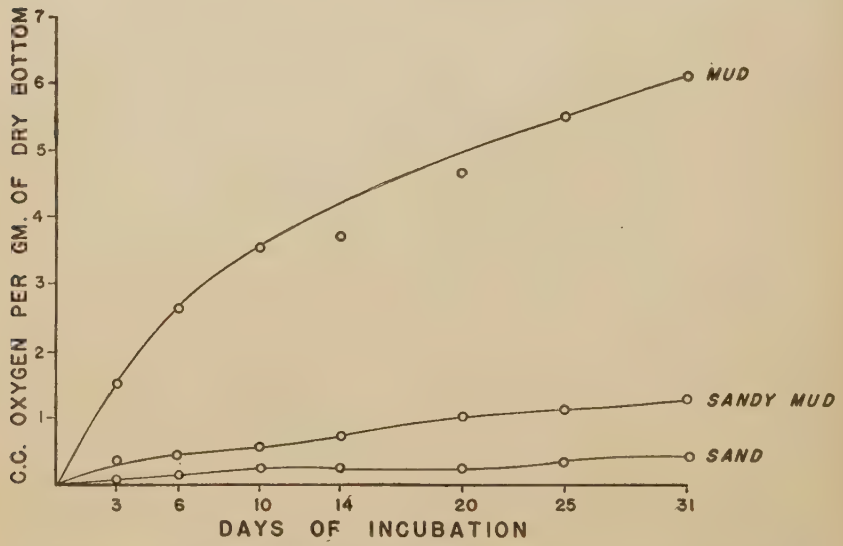

Figure 34. Course of oxygen consumption by different types of bottom material.

mud bottom containing 1.46% organic matter absorbed 1.31 cc. oxygen per gram; while the mud bottom with 5.82% organic matter absorbed 6.14 cc. oxygen per gram. The increases in oxygen consumption were greater than the increases in the organic matter content of the bottom material. This is shown by the ratio of oxygen in cc. to organic matter in mg., which was found to be 0.055 for the sand, 0.09 for the sandy mud, and 0.106 for the mud. The higher value for mud bottom points to the greater availability of its organic matter. Thus the degree of oxidation seems to depend on the presence of biological agents and the nature of the organic matter.

Some idea of the effect of different types of organic matter may be gained from a report on the experiments on oxygen absorption by material from different portions of the same core. Fresh cores of mud were obtained about

4 miles west by north of Gay Head. These cores were divided into 3 sections: 0–1 cm., 1–5 cm., and 5–20 cm. The moisture content of these sections of mud was 45.8, 33.7, and 26.7 per cent, respectively. One gram portions of the surface layer and 2 gm. portions of the lower layers of the fresh mud were added to a series of oxygen bottles; fresh sea water was used and the bottles were incubated for three days. The results presented in the following summary show that the surface layer of mud was somewhat more active biologically than the deeper layers, and this layer had the higher organic content.

Depth of layer cm.	*Mud used, dry basis gm.*	*Oxygen absorbed in 3 days cc.*	*Oxygen absorbed per 1 gm. of dry mud cc.*	*Nitrogen content of dry mud per cent*
0–1	0.54	0.17	0.315	0.095
1–5	1.32	0.35	0.265	0.087
5–20	1.47	0.40	0.272	0.070

In order to establish the relation of bacterial development to the oxidation of organic matter in the mud, the results of the following experiment may be cited. A mud core obtained in Buzzards Bay, near Weepecket Buoy, was divided into 3 sections. The upper layer of mud was suspended in some of the surface water and 5 cc. portions pipetted into a series of oxygen bottles; in the case of the other two layers 2 gm. portions of the moist material were used. The sea water menstruum was 24 hours old. There was a marked increase in bacterial activities in the water receiving marine mud. The surface layer of the mud was again the most active biologically. (Table III.)

TABLE III

BACTERIAL MULTIPLICATION AND OXYGEN CONSUMPTION OF DIFFERENT LAYERS OF A MARINE MUD PROFILE

Material used[1]	*0 days*		*1 day*		*5 days*		*Oxygen consumed per 1 gm. of dry mud*
	Bacteria in 1 cc.	*Oxygen, cc. per bottle*	*Bacteria in 1 cc.*	*Oxygen, cc. per bottle*	*Bacteria in 1 cc.*	*Oxygen, cc. per bottle*	
Standard sea water control	9,000	1.11	30,500	1.08	1,400	1.00	—
Standard sea water + 25 gm. dry mud 0–2 cm. deep[1]	4,000	0.95	83,000	0.74	19,800	0.49	1.84
Standard sea water + 1.05 gm. dry mud 2–6 cm. deep	7,500	0.80	114,500	0.68	15,800	0.31	0.47
Standard sea water + 1.05 gm. dry mud 6–10 cm. deep	5,500	0.85	116,000	0.69	9,700	0.19	0.63

[1] Nitrogen content of dry material, 0.146% for 0–2 cm., 0.180% for 2–6 cm., and 0.138% for 6–10 cm. layer.

A detailed study was made of the biological oxygen absorption by the A and B layers of a freshly taken mud profile. The position of the Station was off Gay Head. The A layer comprised the upper 2 cm. of mud and the B layer the lower 3 cm. (2–5 cm. depth). The nitrogen content of the mud was, on a dry basis, 0.176% for the A layer and 0.063 for the B layer. Cultured sea water was used and the oxygen determinations made in siphoned water. Different amounts of mud were added to the bottles. The curves presented in Fig. 35 are based upon the averages of the results obtained on the basis of 1 gm. portion of dry mud. The results again prove that the upper layer of mud exerts a greater oxygen absorption than the

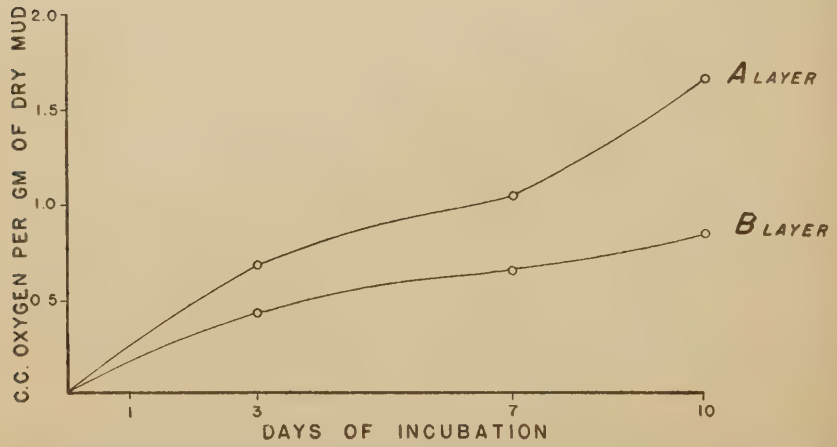

Figure 35. Course of oxygen absorption by surface (A) and sub-surface (B) layers of a Gay Head mud core.

lower layer and that the process, although slow, is gradual and continuous. The greater oxygen absorption by the mud from the upper layer may be due, in this case as well, to the greater organic matter content.

In order to demonstrate further the biological nature of the oxidation process taking place in marine bottom material added to sea water, a study was made of the influence of temperature upon oxygen consumption. Cultured water, 3 weeks old, and fresh mud taken off the coast of Cape Cod, in line of Provincetown were used. This mud contained 0.234% nitrogen, on a dry basis. Five cc. portions of a suspension of mud in sea water, equivalent to 0.26 gm. of dry mud, were added to a series of oxygen bottles. These were filled with cultured sea water and incubated, under water, at 3 different temperatures: the lowest, or refrigerator temperature, ranged, after the first 24 hours, from 2° to 6° C., with an average of 4° C.;

the medium or room temperature, ranged from 21° C. to 25° C., average 23° C.; the highest, or incubator temperature, was quite constant, after 48 hours, ranging from 31.0° C. to 31.9° C., averaging 31.5° C.

The results (Table IV) show the increase in oxygen consumption effected by the higher temperature of incubation, at which temperature, there was comparatively little oxygen consumed after 10 days and very few bacteria left in the water after 15 days. At the low temperature, there was an extensive population of bacteria in the water containing the mud, but these bacteria had a low oxygen absorbing capacity.

TABLE IV

Influence of Temperature upon the Oxygen Consumption of Marine Mud

Temperature of incubation	*Period of incubation*	*Oxygen consumed, cc.*			*Bacteria in 1 cc. water (+ mud)*
		per bottle		*per 1 gm. of dry mud*	
		water alone	*water + mud*		
°C.	*days*				
4° C.	5	0	0.05	0.19	528,000
4° C.	10	—	0.15	0.48	—
4° C.	15	0.05	0.21	0.62	49,200
23° C.	5	0.04	0.24	0.77	182,000
23° C.	10	—	0.34	1.00	—
23° C.	15	0.12	0.47	1.35	4,200
31.5° C.	5	0.10	0.33	0.89	382,000
31.5° C.	10	—	0.49	1.49	—
31.5° C.	15	0.11	0.52	1.58	<100

It was found in this, as well as in other experiments, that when a rise in bacterial numbers accompanying active oxidation of organic matter in water or in bottom material is followed by a rapid drop in numbers, oxygen consumption continues, although at a slower rate, in the presence of the few bacteria capable of developing on an agar plate. Similar observations have been made by Allgeier, Peterson, Juday and Birge (*1*) who found, in the case of lake sediments, that after gas evolution had reached a maximum, during the anaerobic decomposition of these sediments, bacterial numbers diminished rapidly; this drop was not accompanied by any marked decrease in gas evolution. This continued oxidation, after bacterial numbers have been markedly reduced, may be due to the continued action of bacterial enzymes or activating factors, as suggested by Kreps (*3*), who emphasized that bottom deposits are particularly rich in such enzymes. It may also be due to the controlling effect upon the bacterial population by the animal population in the water, as suggested elsewhere (*13*). Of interest in this connection is the work of Butterfield (*6*) who found that in the presence of certain protozoa, a decline in bacteria took place, not accompanied, however, by a diminution in the oxygen consumption in the water.

What is the greatest oxygen consumption which can be obtained with fresh marine mud under optimum laboratory conditions? It seemed possible to determine this if a new supply of aerated water was used to replace the sample removed by siphon for oxygen determination. A series of oxygen bottles received 1.0, 2.5, and 5.0 gm. portions of fresh marine mud. One series of bottles was placed in the dark under water, in an undisturbed state, while the bottles in the other series were shaken for a half minute, by hand, several times daily. At intervals of 1 to 4 days, the supernatent clear water was siphoned off and analyzed for oxygen. The bottles were immediately refilled with fresh cultured water, which had been saturated

TABLE V

MAXIMUM OXYGEN CONSUMPTION OF FRESH MARINE MUD UNDISTURBED AND RE-SUSPENDED

Incubation days	*Oxygen consumed per bottle*						*Bacteria in 1 cc. of water*[3]	
	Bottles shaken			*Bottles unshaken*				
Amount of mud used per bottle[1]	*1 gm.*	*2.5 gm.*	*5 gm.*	*1 gm.*	*2.5 gm.*	*5 gm.*	*0*	*1 gm.*
1	0.53	0.79	1.08	0.30	0.43	0.51	51,200	279,000
3	0.51	0.76	0.95	0.29	0.32	0.32	51,900	89,500
5	0.24	0.49	0.57	0.31	0.48	0.49	—	—
10	0.42	1.12	1.30	0.64	1.26	1.51	56,900	161,000
15	0.14	0.65	0.74	0.28	0.70	0.88	—	—
22	0.18	1.08	1.09	0.27	1.01	1.51	12,500	12,000
30	0.03[2]	0.60	1.41	0.09	0.87	1.55		
Total	2.05[2]	5.49	7.14	2.18	5.09	6.77		

[1] Moisture content of fresh mud 47%.
[2] 26 days incubation.
[3] In supernatant water above the mud.

with oxygen, and incubated longer. The mud was obtained from Buzzards Bay, near Weepecket and contained 47 per cent moisture. The nitrogen content of the dry mud was 0.210%. The results reported in Table V represent the differences between the water controls and the water receiving the different amounts of mud. The total oxygen consumption per 1 gm. of dry mud, for a period of 28–30 days, was found to decrease with the amount of mud used; it amounted to 4.0 cc. when 1 gm. of moist mud was used per bottle; 3.99 cc. for 2.5 gm. mud, and 2.62 cc. for 5.0 gm. mud. Although the mud exerted a certain adsorptive effect upon the bacteria in the water, the numbers of bacteria were usually much higher in the water with mud than in the water alone.

A detailed study was now made of the oxidation of various types of bottom material collected from different depths of the sea and at different distances from land. Usually not over 7 days elapsed before taking the

samples and subjecting them to oxidation in cultured sea water. Soon after the cores were obtained, they were placed, in an undisturbed position on ice. When brought to the laboratory, the cores were removed aseptically from the glass tubes, and divided into two portions: the upper 1 to 4 cm. or A layer and the lower or B layer, the latter varying from 4 to 25 cm. in thickness. Aliquot portions of the bottom material were placed in oxygen bottles; these were filled with cultured water, 7–15 days old, and the bottles incubated at room temperature in the dark and under water. Usually 5 bottles were employed for each sample: one bottle was analyzed immediately after the sediment had settled for 1–2 hours; 2 bottles were analyzed after

TABLE VI

OXYGEN CONSUMPTION BY BOTTOM MATERIAL IN THE NEIGHBORHOOD OF WOODS HOLE

On basis of dry material

Sta. No.	*Position of station*	*Depth of station*	*Layer*	*Nitrogen*	*Oxygen consumed, cc. per 1 gm. in 4 days*
		meters		%	
1	Gay Head, 5¼ mile W by S	36.0	A	0.086	0.42
		36.0	B	0.070	0.29
2	Gay Head, ½ mile SE of Sta. 1	38.0	A	0.080	—
		38.0	B	0.067	0.18
3	Buzzards Bay 2 miles WNW of Weepecket Rock	15.7	A	0.257	1.10
		15.7	B	0.221	0.80
4	Same as 3, 1 mile NWW of Weepecket Rock	16.0	A	0.227	1.14
		16.0	B	0.182	0.93
5	Same as 3, 1/3 mile N of Weepecket Rock	15.3	A	0.255	0.82
		15.3	B	0.222	0.87
6	Same as 3, 1 mile NE of Gong 6	16.0	A	0.259	1.04
		16.0	B	0.222	0.94
7	Same as 3, Buoy 4LB	37.0	A	0.085	0.13
		37.0	B	0.048	0.05

4–7 days and the other two after 8–15 days; only the results of the final analyses are reported in the tables.

In the first experiment (Table VI), a group of bottom samples were collected from stations in the neighborhood of Woods Hole; the water was 7 days old, and the oxidation period only 4 days. In making the calculations, allowance was made for the chemical oxidation or oxygen consumption within 1 to 2 hours, and oxygen consumption by the water controls. The results reported in this and other tables thus represent the biological oxygen consumption due to the oxidation of the bottom material in the water; they are reported on the basis of 1 gm. of dry material. A marked correlation was found between oxygen consumption and organic matter content of the bottom material. The cores from stations 3–6 are higher in organic matter and show also more active oxygen consumption. By comparing the

average organic matter content of these stations with those containing less organic matter, using the results for the two layers, one finds that the organic matter in the former oxidizes more readily than in the latter.

Station Nos.	*Organic matter content* %	*Oxygen consumed, cc. per liter*	*cc. of oxygen / mg. of organic matter*
3–6	4.50	0.95	0.0210
1, 2, 7	1.45	0.21	0.0145

In the following two experiments, bottom samples were obtained at varying distances from land. In the first Atlantis cruise, the depths of the stations ranged from 104 meters to 1,250 meters. With the exception of the shallow station, coming from a depth of 104 meters, the organic matter from these bottom samples showed comparatively little oxygen consumption (Table VII). Although the organic matter content of these bottom samples was about 2 to 3 per cent, as calculated from the total nitrogen, there was only 0.07–0.27 cc. oxygen consumed per 1 gm. of dry material, in 14 days. These results suggest that the organic matter in the sea bottom at greater distances from land has a considerable resistance to decomposition. This was found to hold true of the oxidation of the organic matter in the bottom samples obtained on the second cruise of the Atlantis, when the samples came from even greater depths. The highest oxidation was 0.59 cc. per 1 gm. of dry bottom material, although this material contained 2.5 per cent organic matter. These results thus point definitely to the lower oxidizability of the organic matter in the deep sea bottoms.

A comparison of the data for the deeper stations, presented in Tables VII and VIII with those of the shallower stations (Table VI) points to another interesting difference: in the case of the bottom samples close to land, the upper layer of the bottom material gave greater oxygen consumption than the lower layer; however, in the case of the deeper stations (7, 8, 9 in Table VIII), the lower layer of bottom material frequently showed as great if not greater oxygen consumption than the upper layer.

In order to study further the oxidizability of the organic matter in sea bottom not far distant from land 3 groups of stations were selected as representing characteristic differences in depth and position. One group (stations 1–8) was obtained in line of the Isles of Shoals; the other group (stations 9–13) came from the 99 fathom hole in Massachusetts Bay, and the third group (stations 14–17) from a line off Provincetown. The results, presented in Table IX, show that the consumption of oxygen or the decomposibility of these bottom samples was greater than in the case of the bottom material which had been obtained at greater distances from land. There was also a certain definite correlation between the rate of oxidation and the depth or position of the bottom material which is brought out in

the group of samples from the 99 fathom hole (stations 9–13). In this hole the deeper stations gave higher organic matter content and the oxygen consumption by the bottom material was greater. A certain correlation

TABLE VII

OXYGEN CONSUMPTION BY BOTTOM SAMPLES FROM FIRST "ATLANTIS" CRUISE
On basis of dry material

Sta. No.	*Position of station*	*Depth of station*	*Layer*	*Nitrogen*	*Oxygen consumed, cc. per 1 gm. in 14 days*
	W Lat. N Long.	*meters*		%	
1	40.04′ x 70.50′	104	A	0.191	0.63
		104	B	0.154	0.65
2	39.56′ x 70.49′	175	A	0.046	0.10
		175	B	0.048	0.07
3	39.52′ x 70.48′	600	A	0.105	0.13
		600	B	0.087	0.15
4	39.50′ x 70.46′	914	A	0.106	0.07
		914	B	0.102	0.08
5	39.48′ x 70.47′	1,250	A	0.154	0.27
		1,250	B	0.139	0.17

TABLE VIII

OXYGEN CONSUMPTION BY BOTTOM SAMPLES OBTAINED IN SECOND "ATLANTIS" CRUISE
On basis of dry material

Sta. No.	*Position of station*	*Depth of station*	*Layer*	*Nitrogen*	*Oxygen consumed, cc. per 1 gm. in 15 days*
	W Lat. N Long.	*meters*		%	
6	40.14′ x 67.46′	1,134	A	0.030	0.31
		1,134	B	0.040	0.08
7	40.13′ x 67.39′	1,609	A	0.130	0.26
		1,609	B	0.112	0.41
8	40.09′ x 67.23′	2,100	A	0.133	0.39
		2,100	B	0.124	0.59
9	40.04′ x 66.67′	2,967	A	0.118	0.27
		2,967	B	0.089	0.34
10	40.25′ x 66.00′	4,000	B	0.034	0.11

was also found between the amount of organic matter in the bottom material and the rate of its oxidation.

In all the above experiments, the water was tested, at the end of the oxidation period, for ammonia, nitrite and nitrate.[1] In no case could these be detected save in mere traces. This might have been due to the fact that the incubation period of 14–15 days was not sufficient for the liberation of the nitrogen in appreciable or detectable amounts in the decomposition of

[1] The authors are indebted to Dr. Cornelia L. Carey for making these tests.

the mud. However, it emphasizes the fact that the nitrogen in the marine bottom is highly resistant and is not readily mineralized by the bacteria. Once the organic residues have settled to the bottom, the nitrogen has become so closely bound in an organic form as to be practically immune to microbial attack. Quite the opposite occurred in experiments on the decomposition of marine plankton (*5, 14*).

In one experiment (Table V), the mud was analyzed for total nitrogen at the beginning and at the end of the decomposition period. There was an increase, on the basis of the dry mud, from 0.210% nitrogen in the fresh

TABLE IX

OXYGEN CONSUMPTION BY BOTTOM MATERIAL FROM A LINE OF STATIONS OFF MASSACHUSETTS BAY, ISLE OF SHOALS AND CAPE COD

Sta. No.	*Position*	*Depth*	*Layer*	*Nitrogen*	*Oxygen consumed, cc. per 1 gm. in 14 days*
	W Lat. N Long.	*meters*		%	
1	42°57′ x 70°31′	83	A	0.193	1.54
		83	B	0.139	0.94
2	42°57′ x 70°27′	100	A	0.209	1.55
		100	B	0.173	1.49
3	42°57′ x 70°25′	100	A	0.138	0.97
		100	B	0.048	0.46
4	42°57′ x 70°22′	132	A	0.252	1.19
		132	B	0.192	2.02
5	42°57′ x 70°20′	157	A	0.291	1.23
		157	B	0.254	1.36
6	42°57′ x 70°18′	161	A	0.306	1.00
		161	B	0.272	1.24
7[1]	42°59′ x 70°15′	155	A	0.288	1.59
		155	B	0.227	1.32
8[2]	42°59′ x 70°12′	175	A	0.326	1.86
		175	B	0.278	1.06
9[3]	42°30′ x 70°18′	90	A	0.081	0.06
10	42°30′ x 70°18′	145	A	0.179	0.87
		145	B	0.051	0.47
11	42°30′ x 70°18′	170	A	0.162	0.83
		170	B	0.176	1.88
12	42°30′ x 70°18′	170	A	0.245	1.33
		170	B	0.211	1.45
13	42°30′ x 70°18′	175	A	0.245	1.71
		175	B	0.228	0.94
14	42°00′ x 69°46′	150	A	0.081	0.62
		150	B	0.047	0.72
15	41°59′ x 69°41′	195	A	0.260	1.41
		195	B	0.261	1.43
16[4]	42°01′ x 69°38′	210	A	0.233	1.11
		210	B	0.234	1.46
17[5]	42°01′ x 69°43′	165–170	A	0.226	1.04
		165–170	B	0.204	1.23

[1] Atlantis Station 2648. [2] Atlantis Station 2654.
[3] Stations 9–13 are from 99 fathom hole in Massachusetts Bay and close to it.
[4] Atlantis Sta. 2645. [5] Atlantis Sta. 2646.

mud to 0.233% nitrogen[1] in the mud left after 30 days' oxidation. If one is to assume that this increase in the nitrogen content of the mud is accompanied by a narrowing in the carbon-nitrogen ratio and, therefore, by a corresponding decrease in the carbon content, the amount of carbon lost by oxidation would correspond roughly to 2.3 mg. carbon per 1 gm. of dry mud oxidized. The amount of oxygen theoretically required for the oxidation of this carbon would be 4.29 cc. The actual oxygen consumption varied from 2.62 to 4.00 cc. oxygen, depending on the amount of mud inoculum, as shown above. This brings added proof that the nitrogen was not liberated in a mineralized form since it could be accounted for almost quantitatively in the residual mud.

DISCUSSION

A knowledge of the abundance, chemical nature and rapidity of decomposition of the organic matter in the sea bottom represents a number of significant theoretical and practical problems. Among these, the question of the influence of the nature of the organic matter upon the occurrence of bottom-feeding or benthotic animals, as well as the decomposition of the organic matter by bacteria and the return of the elements into circulation are not the least important. To what extent does the content of organic matter in the sea bottom influence the abundance of bottom feeding animals? Is the occurrence of the latter connected with the amount of organic matter or with the nature of this organic matter? Is the organic matter subject to decomposition or oxidation by bacteria, and how can this best be measured?

The determination of the organic matter content in the sea bottom represents little difficulty. It can be done roughly by measuring the loss on ignition, but more accurately by an analysis of the organic carbon or the total nitrogen content. Since the ratio between these two elements is more or less constant in the marine organic matter or marine humus, the determination of one of these two elements may be sufficient as a measure of the abundance of humus. The carbon can be multiplied roughly by 2 and nitrogen by 20, to give the total organic matter in the bottom material. A knowledge of the nature of the organic matter represents, however, a much more difficult problem. Chemical analyses, as attempted by one of us (*11*) or hydrolysis by means of proteolytic enzymes, as attempted by Boysen-Jensen (*4*), are either too complicated or are not comparable with marine processes. The measurement of the rapidity of oxidation of the bottom material, under conditions similar to those taking place in nature, namely in a sea water system with favorable oxygen tension, offers a much simpler method for comparing different types of marine bottom material. The oxidation of the organic matter in sea water was found to be a convenient

[1] The 3 samples of mud contained 0.230%, 0.238%, and 0.232% nitrogen.

method for measuring the relative availability or stability of this organic matter.

Three of the stations from which material was collected for the present investigation have also been used the year previously in a study made by Dr. Hjort and Dr. Bigelow concerning the relation of sea bottom to the occurrence of shrimps. It is of interest to compare, for these 3 stations at least, the abundance of organic matter, the rate of oxidation and the occurrence of shrimps:

Sta. No.	*Shrimp catch (1936)*	*Nitrogen content of mud bottom, dry basis, per cent*		*Oxygen consumption cc./1 gm. in 14 days*
		(1936)	*(1937)*	
	qts.	*A layer*		
2645	1.5	0.217	0.233	1.11
2648	16.0	0.295	—	1.59
2654	180.0	0.309	0.326	1.86

The above results show a very interesting correlation between the occurrence of shrimps and the rapidity of oxidation of the organic matter in the bottom material. The number of samples used for this comparison is too small, to justify any broad conclusions; but the results are highly suggestive.

Of particular interest in connection with these studies is the fact that the organic matter in the sea bottom is oxidized only very slowly by bacteria. It was found that 1 mg. of organic matter consumed only 0.1 cc. oxygen, when allowed to oxidize, under optimum conditions in the laboratory, for a period of 15 to 30 days. Assuming 50 per cent carbon in the organic matter, only about 10 per cent is found to become oxidized in this period of time. When this is compared with the rapid decomposition of the plankton material (*5*, *14*), the differences are particularly striking. This points to great resistance of the organic matter in the sea bottom to decomposition.

Even more important than the limited decomposition of the organic matter is the fact that the nitrogen is not mineralized at all in the period of time used, namely in 15–30 days. The oxidation process is thus found to take place at the expense of the non-nitrogenous organic complexes in the bottom material. The fact that this organic matter is not uniform in composition, but is made up of a number of organic compounds has been demonstrated previously (*11*).

Of further significance is the fact that the rate of oxidation of the organic matter from bottom deposits nearer to land and from shallower basins is considerably greater than that from bottoms at greater depths and distances. The organic matter in these bottoms is much more resistant, the oxidizability being only one-tenth or one-fifth of that of the organic matter in shallower bottoms near land. This suggests that the organic matter of the latter is more "aged" or has had time to become more highly oxidized.

SUMMARY

1. A study has been made of the rate of oxidation of the organic matter in the sea bottom, as influenced by its abundance, nature of bottom material, distance from land and depth of bottom.

2. A convenient procedure was developed for this study. It consisted in placing definite amounts of fresh bottom material in oxygen bottles containing a known volume of "cultured" water (water kept for several days in the laboratory and resaturated with oxygen). The bottles were incubated for 7–15 days or longer; the clear water was then siphoned into smaller oxygen bottles and the amount of oxygen measured. The results have been calculated on the basis of 1 gm. of dry material. Comparisons were made between the organic matter content, as calculated from the total carbon or nitrogen in the material, and the amount of oxygen consumed.

3. Sandy bottoms were found to consume very little oxygen chemically, while mud bottoms consumed larger amounts. The biological consumption of oxygen by different types of bottom material was considerably greater than chemical absorption.

4. By allowing oxidation of the organic matter to proceed for 15–30 days, it was found that 1 milligram of organic matter in bottoms close to land would consume about 0.1 cc. of oxygen, under the most favorable conditions.

5. The organic matter in sea bottom at greater depths was oxidized to a much lesser extent than the organic matter in shallower bottoms, nearer to land.

6. The organic matter in the sea bottom was found to be much more resistant to biological decomposition or oxidation than had been previously determined for the organic matter in sea water or in the plankton organisms.

7. The oxidation of the organic matter, at least for 15–30 day periods, took place at the expense of the non-nitrogenous constituents; very little of the nitrogen was liberated in a mineralized form in that period of time.

8. It is suggested that the rate of oxidation of the organic matter in the bottom material be used as a measure of the availability or nature of this material.

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ON THE VALIDITY OF THE DYNAMIC TOPOGRAPHIC METHOD FOR THE DETERMINATION OF OCEAN CURRENT TRAJECTORIES

By

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In a previously published statement (Parr 1937, b) the writer has listed the main reasons why, in the present state of our knowledge of the actual as distinct from the ideal dynamics of ocean currents, he feels inclined to take greater confidence in the evidence of the actual paths of flow obtained from a study of the identifying properties of the water masses, than in the conclusions reached by present efforts to deduce the trajectories of motion from hydrodynamic calculations and considerations. In this article a further elaboration and elucidation of these objections against the unrestricted use of dynamic methods for the determination of ocean current trajectories will be attempted, while suggestion for obtaining a presumably closer approach to an accurate presentation of flow through isopycnic analysis by means of identifying properties will be presented in the next article in this issue.

It should be clearly understood, however, that the objections against the dynamic methods here advanced are not directed against the theoretical validity of Bjerkness' circulation theorem under ideal abstract circumstances, but against the practical application of this theorem to actual conditions in nature, on the working hypothesis that the direction of flow will always be at right angles to the dynamic gradient (or parallel with the dynamic contour lines) with general disregard from the question of whether these actual conditions adequately fulfill the theoretical requirements for a valid application of the method.

With reference to the dangers inherent in the unrestricted use of dynamic topographies on the basis of the practical working hypothesis above mentioned, the writer has already summarized his objections in the statement "that the dynamic methods of analysis deal only with states of motion and not with the trajectories or paths of transportation, and, since the error in the calculations for states of motion is cumulative with reference to trajectories, an accuracy which is adequate for approximate determination of states of motion may be utterly inadequate for analysis of actual paths of transportation" (Parr, 1937 b). In the following we shall consider the reasons why it seems probable that it will prove the exception rather than the rule under natural conditions to find the ideal requirements for the

dynamic method sufficiently fulfilled to make the accuracy of this method adequate also for the study of current trajectories.

The first and most general difficulty arises from the fact that dynamic topographies can only deal with components of motion in the two horizontal dimensions, but are incapable of including the third dimension in their presentation of the probable current picture. In this simple fact, we find an objection of universal character, since all liquid flow is basically three-dimensional and can never find full expression in a two-dimensional presentation except in the form of a rough approximation where the vertical components are negligible. If we nevertheless arbitrarily select to assume that the contour lines are indeed identical with the actual lines of flow, we not only *disregard* the probable existence of vertical components of motion, but actually by our assumption *rule out* the possibility of any vertical movement whatsoever (except by the process of diffusion), since the dynamic contour lines must always form closed curves* therefore presenting for each isobaric surface a completely self-contained and self-compensating circulation system, not permitting of any invasion from or evasion to other isobaric surfaces. True convergence or true divergence can therefore never become apparent in the dynamic topographies alone, and the cases in which the dynamic interpretation becomes most obviously unsatisfactory are therefore the cases in which these phenomena are definitely known to occur.

While the successful prediction of the movement of icebergs accomplished by the U. S. Ice Patrol from a study of dynamic topographies is one of the greatest practical achievement of applied oceanography, of immeasurable value to human lives, it must be remembered that unless the liquid flow itself is strictly confined to two dimensions only, the movements of a rigid body floating in, or on, a complicated liquid current system will only express a mechanical integration of the horizontal components of the liquid flow upon which the object is carried, without any direct relation to the actual liquid trajectories. Comparison might be made with the movements of an object transported on a conveyor system of rollers rotating in the same direction. The movement of the object would express the mechanical integration of direction and velocities of the movement of the upper roller surfaces reduced to two dimensions, but could not be taken as an indication of the nature of the roller motion itself, unless its mechanism was known to us from entirely different sources of information.

Both because the conditions at the meeting of the Labrador Current and the so-called Gulf Stream in the vicinity of the Newfoundland banks presents us with perhaps the most unquestionable case of a true convergence, undoubtedly at least in considerable part involving movements in three

* Apart from the definitely *not* ideal cases to be dealt with later, in which the dynamic contour lines have their end point or their point of origin at the shore.

dimensions, and also for the very reason of the successful application of the dynamic topographic method to the *approximate* prediction of the mechanical integration, in the movements of the icebergs, of the two-dimensional components of this three-dimensional flow; the hydrographic observations recorded by the U. S. Ice Patrol may provide us with a particularly suitable example for a study of the relationship between actual liquid trajectories and the dynamic contour lines, when movement occurs in more than two dimensions.

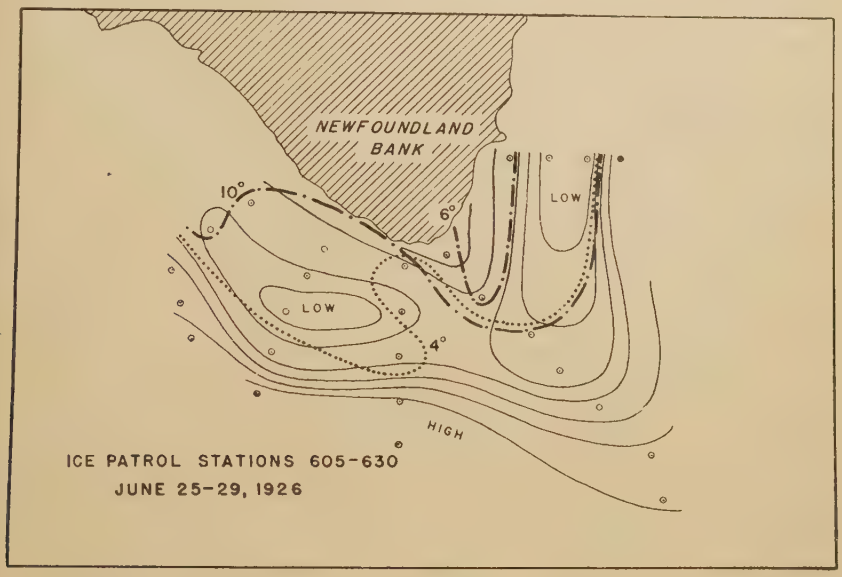

Figure 36. Surface isotherms for 10°C. and 6°C. (——·——·) and intermediate isotherm for 4°C. (·····) superimposed upon dynamic topography of free surface relative to 750 decibars, according to U. S. Ice Patrol observations for June 25-29, 1926 (Smith, 1927).

The report on the international ice observations for the season of 1926 (Smith 1927) gives the data on which the accompanying Fig. 36 is based. In this figure, representing conditions during June 25–29, 1926, the dynamic contours are drawn in light solid lines according to the interpretation of Smith (1927, Fig. 57, p. 115). Superimposed upon this dynamic topography in heavy dot and dash are the surface isotherms for 10° C. and 6° C., and in heavy dotted lines the southern limits of occurrence of water colder than 4° C. as an *intermediate* layer below the surface. In this picture we find some agreement in broad generalities between the distribution of surface temperatures and the trajectories indicated by the contour lines, but a glaring discrepancy in the details, particularly when we notice how the contour

lines coming in from the north with cold surface waters of northern origin having temperatures of less than 6° C. and salinities of less than 33‰, depart again to the north in what would purport to be a continuous trajectory carrying warm surface waters of unquestionably southern derivation with temperatures of more than 10° C. and salinities of nearly 34‰. On the assumption that contour lines and true trajectories should be identical, this transfer of surface waters should be utterly impossible. But it is evident that surface waters of southern origin must have traversed the contour lines towards the north, that is, they must have descended a dynamic gradient calculated with reference to a fixed datum level, to a considerable extent. If it should be felt that the magnitude of the discrepancy in terms of miles, within the small area illustrated, it not so exceedingly great, it should be kept in mind, first, that the discrepancy must actually be greater than indicated by the isotherms (see below) and, secondly, that even if we assume that actual trajectories, after this region of convergence has been passed, will approximately follow the contour lines in their further course across the North Atlantic, an error of only a few miles from one trajectory to another within this small area of convergence may amount to an error of hundreds of miles in the eastern part of the ocean.

The discrepancy patently arises from the fact that we here have a true convergence with the actual trajectories of the incoming current from the north submerging under the surface trajectories of the supernatant current from the east and south. Since the dynamic contour lines do, and as above mentioned always must, enter and leave the picture in equal numbers, we here have an excellent illustration of the failure of the contour lines to indicate true trajectories in the presence of significant vertical components.

That we are actually dealing with a true submergence may be seen by the extent of the southward distribution of the *intermediate* layer colder than 4° C. far beyond the northward distribution of the surface waters of southern origin warmer than 10° C. In making this comparison, one must keep in mind that the illustration undoubtedly greatly minimizes the extent of the overlap between southern and northern waters, due to the fact that both limiting isotherms have been pushed back towards their region of origin in relation to the actual extent of the mutual invasions which gave rise to their distribution; the limiting isotherm for surface waters warmer than 10° C. by chilling from the atmosphere and from the waters below, that of the intermediate water colder than 4° C. by warming both from above and from below.

The actual existence of this convergence and submergence of northern waters is, of course, already well known and recognized by all investigators of North Atlantic circulation, and it can be even better illustrated than in Fig. 36 by the observations made by the Ice Patrol somewhat earlier in the same season (May 14–20, 1926, Smith, 1927). These observations are plotted in Fig. 37, in which the surface isotherm of 4° C., in heavy solid line,

and the limits of southern subsurface penetration of waters colder than 4° C. at intermediate levels, in heavy dotted line, are superimposed upon the dynamic topography of the surface. In this case the descent of water colder than 4° C. from the free surface has occurred within the region of observation, and the continuity of the descent can be followed. If we make a composite profile through the region of greatest southward penetration of this intermediate submerged tongue of northern water, by projecting the sta-

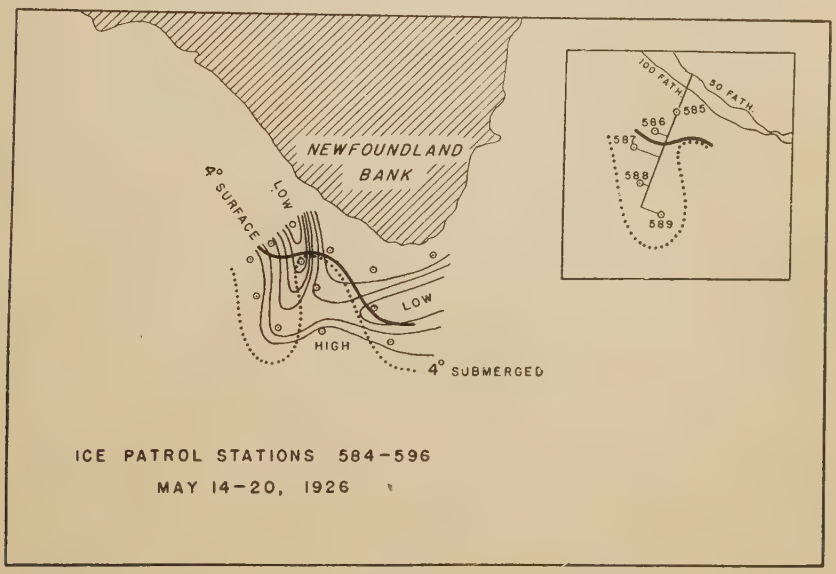

Figure 37. Surface isotherm for 4°C. (————) compared with isotherm for 4°C. at intermediate depths (·····) superimposed upon dynamic topography of free surface relative to 750 decibars according to U. S. Ice Patrol observations for May 14–20, 1926 (Smith, 1927). Insert: arrangement of composite profile shown in fig. 38.

tions within this tongue to a straight line as shown in the insert in Fig. 37, we obtain the picture given in Fig. 38 (with the individual station curves for temperature shown at the right). From this dichotomous picture it is rather clearly indicated that both types of submergence to be expected in the neighborhood of melting ice must have occurred in the vicinity of Newfoundland Bank, namely an almost vertical convection of bottom water, and an oblique submergence of intermediate water, both of northern origin. This, one will notice, corresponds very closely and interestingly to conditions in the Antarctic (see Wüst, 1928; Sverdrup, 1933), although on a much smaller scale.

From these pictures it is evident that there must be an oblique move-

ment from the surface downwards and southwards, and this invasion of heavier water from the north at intermediate levels must cause a decrease in the dynamic elevation of the surface calculated with reference to a fixed datum level, which will be at variance with the actual northward extension of the surface trajectories.

When the Ice Patrol reports repeatedly mention cases in which the icebergs actually seemed to move almost at right angles to the dynamic contour lines, this is undoubtedly often, at least in part, due to the fact that the more deeply submerged portions of the berg are still being forced southward

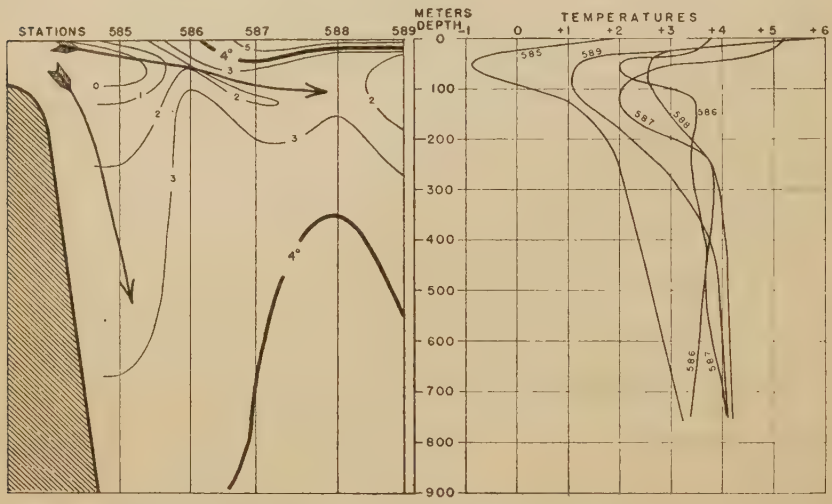

Figure 38. Composite profile off southern edge of Newfoundland bank (see insert in fig. 37) according to U. S. Ice Patrol observations, May 14–20, 1926 (Smith, 1927). Temperature curves for individual stations at right.

by the submerging Arctic waters. In one instance when such transverse movements are reported (Stations 1109–1111, International Ice Observations, Season of 1930) two of the three stations in the series concerned actually show a strong density inversion associated with the intermediate submergence of northern waters, each inversion confirmed by two observations in the same vertical. This is a condition which can only exist as a transitory state in the presence of high velocities of transverse movement in the intermediate cold layer.

On the whole, conditions at the southern end of the Newfoundland Bank would thus seem to furnish an excellent example of the inadequacy of dynamic topographies for the interpretation of current trajectories in regions of true convergence.

A further difficulty which may become important in special cases, but which is of less general significance, is caused by the influence of centrifugal forces whenever there is a change in the radius of curvature of the current trajectories. As previously stated (Parr 1937b) strictly ideal contour line flow can even under otherwise ideal circumstances only occur in a straight parallel, or in a concentric circular system. Here again we may find an example in the U. S. Ice Patrol observations. Fig. 39, comparing dynamic topography with recorded movements of icebergs, is taken from Smith (1931, Fig. 106, p. 165) and shows a centrifugal tendency of the bergs in relation to the contour lines with a rapidly changing radius of curvature.

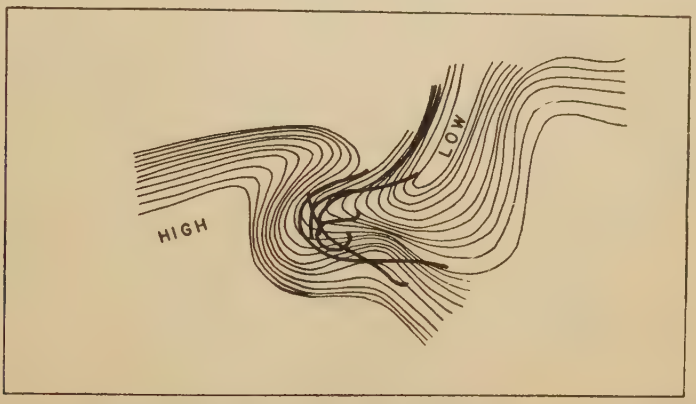

Figure 39. Drift of icebergs (heavy lines) superimposed upon dynamic topography of free surface near Newfoundland bank June 14–July 3, 1922, according to Smith (1931, fig. 106, p. 165).

Since the effect of the transverse movement of submerged water also enters into the situation (see p. 124), it is far from being a clear case, but centrifugal forces must obviously have contributed to some extent. If, in other instances similar abrupt changes of direction occur without true convergence, the moving water masses must themselves be under the influence of similar centrifugal forces.

A third objection to the expectation that actual conditions in nature will normally, and not only exceptionally, fulfill the requirements for an ideal flow parallel with the dynamic contour lines, is to be found in the fact that the assumption of approximately ideal flow does not take into consideration the loss or gain of dynamic height suffered internally in the direction of motion due to changes in specific gravity caused within the moving mass of water itself by heating or cooling, etc. This objection has already previously (Parr 1936a) been elaborated in greater detail with reference to the changes caused by climatic influences operating from the surface, and a possible

method for obtaining approximate corrections for this factor has been suggested. One might add the statement at this point that the effects of changes in density in the moving mass caused by diffusion from adjacent bodies of water should be similar to those arising from external chilling or heating, and perhaps of even greater general significance.

If, as an abstracting sample we consider that a jet-stream is actually, for whatever dynamic reason, being ejected from a warm body of water into a colder body of similar chemistry (or into a colder region), it is obvious that this jet would suffer an internal contraction and a gain in specific density,

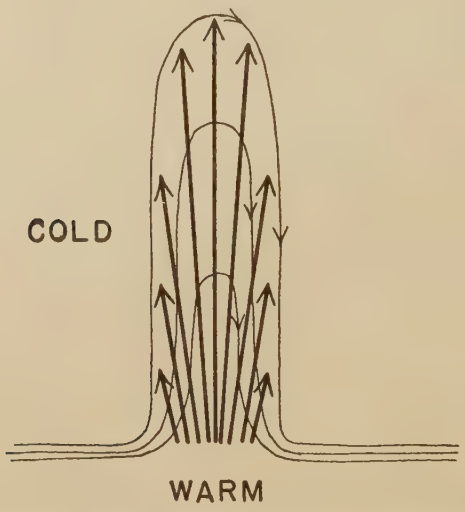

Figure 40. Comparison of contour lines and trajectories in a jet-stream from a warm to a cold region on the northern hemisphere.

with consequent loss of dynamic height, and we would have a situation such as that illustrated in Fig. 40, in which the calculated dynamic contour lines would run at right angles to the actual lines of flow in the middle of the current, and in which the error with regard to direction of motion developing from the assumption of contour-line trajectories in extreme cases might approach 180° on the right side of the current (left side on the southern hemisphere). It seems increasingly evident that the jet-stream type of current (Rossby, 1936) may play a very important role in the general oceanic circulation, and such discrepancies between dynamic contour lines and actual current trajectories as those indicated in this example may therefore have a far more common occurrence than heretofore realized. An apparently excellent case in point, taken from actual observations, may be seen when the dynamic contour lines drawn by Jacobsen (1929, Fig. 41, in

this article) for the region of approach to the British waters is compared with the current map for the North Atlantic prepared by Meyer (1928) from empirical data, and of which a fragment is here reproduced (Fig. 42), with selected contour lines according to Jacobsen entered to facilitate the comparison, which otherwise hardly requires any further elaboration.* Undoubtedly still better illustrations of this particular point could be found in the nearer approaches from the Atlantic to the Norwegian Sea.

The fourth difficulty in the application of present dynamic methods to actual conditions in nature has its basis in the fact that Bjerkness' circulation theorem is valid only in the absence of acceleration or retardation, that is, in the absence of changes in kinetic or potential energy, as explicitly stated by its author (Bjerkness, 1910). The deduction of the velocity equation is also based upon the assumption of absence of friction. It is thus obvious that ideal flow can only occur where and when retardation by friction is exactly balanced by acceleration due to wind or to thermal effects upon the distribution of mass. It seems equally obvious that such a precise equilibrium between accelerating and retarding factors can only be expected to develop in very exceptional cases rather than having their occurrence as the usual event, and that in consequence dynamic contour lines can only under special rather than general circumstances be assumed to parallel the actual current trajectories. In this prerequisite of an absence of change in kinetic or potential energy is further implied that the equations derived from Bjerkness' circulation theorem are strictly applicable only to a parallel system of dynamic contour lines, since, when a convergence or divergence† of dynamic contour lines occurs, the application of the velocity equation will give the result of showing a positive or negative change in velocities and kinetic energy along the same contour line, thereby proving by its own results that the conditions did not fulfill the requirements for a valid application of the method. Although all these reservations were very clearly stated by Bjerkness at the first introduction of the circulation theorem, they would seem to have received only scant attention in general oceanography, and a glance at any dynamic topographic map will give a clear impression of the relatively great rarity of a strictly parallel system fulfilling even this prerequisite for the assumption of ideal flow alone, if not also and simultaneously fulfilling all the other prerequisites as well. At this point it is very important to distinguish again

* The manner in which the Gulf Stream traverses the dynamic contour lines, and its jet-like emergence towards the British waters, is not only confirmed, but even further emphasized by a comparison between Jacobsen's presentation of dynamic topography and Wüst's recent presentation of the current itself (Wüst 1937, fig. 2), which just came to hand after the completion of this article.

† Not to be confused with a true convergence of trajectories with vertical components of motion.

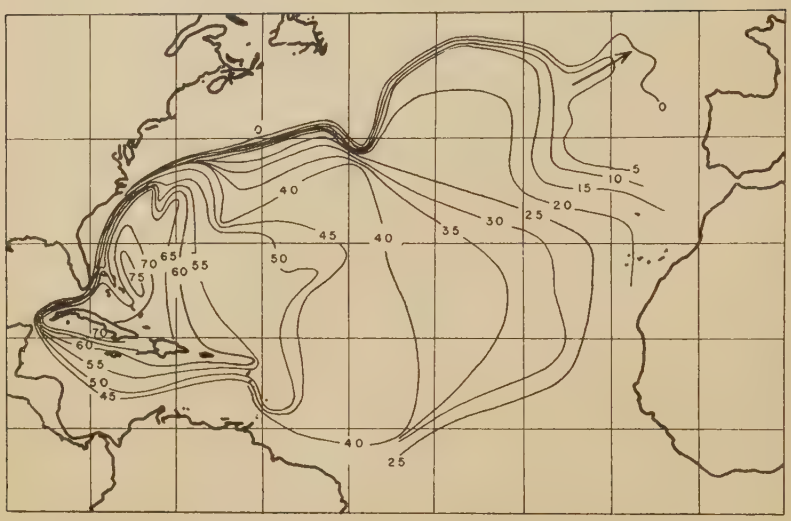

Figure 41. Dynamic topography of the free surface in the North Atlantic according to Jacobsen (1929).

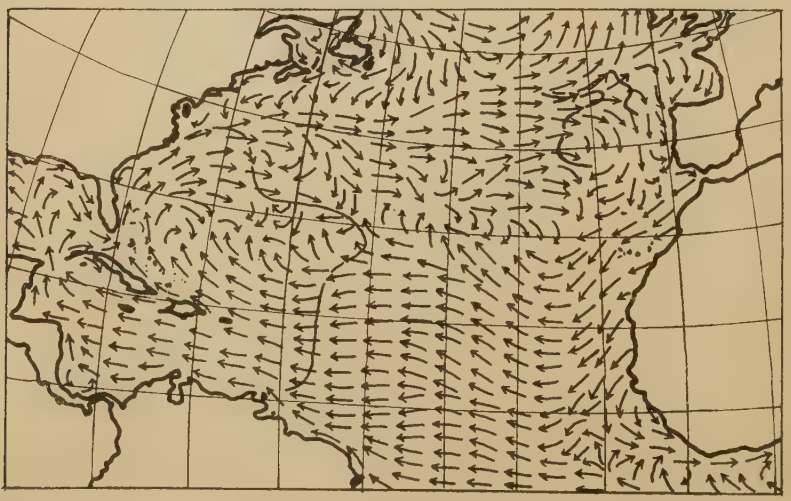

Figure 42. Surface currents in the North Atlantic according to Meyer. Dynamic contour lines of 45, 5 and 0 dyn, cm. relative elevation entered for comparison according to figure 41.

between the approximate determination of states of motion and the effort to trace actual current trajectories, since it seems indicated that if the disturbing influences are only those of retardation or acceleration in the longitudinal direction, without significant interference from any other of the distorting factors discussed in this article, satisfactory results are likely to be

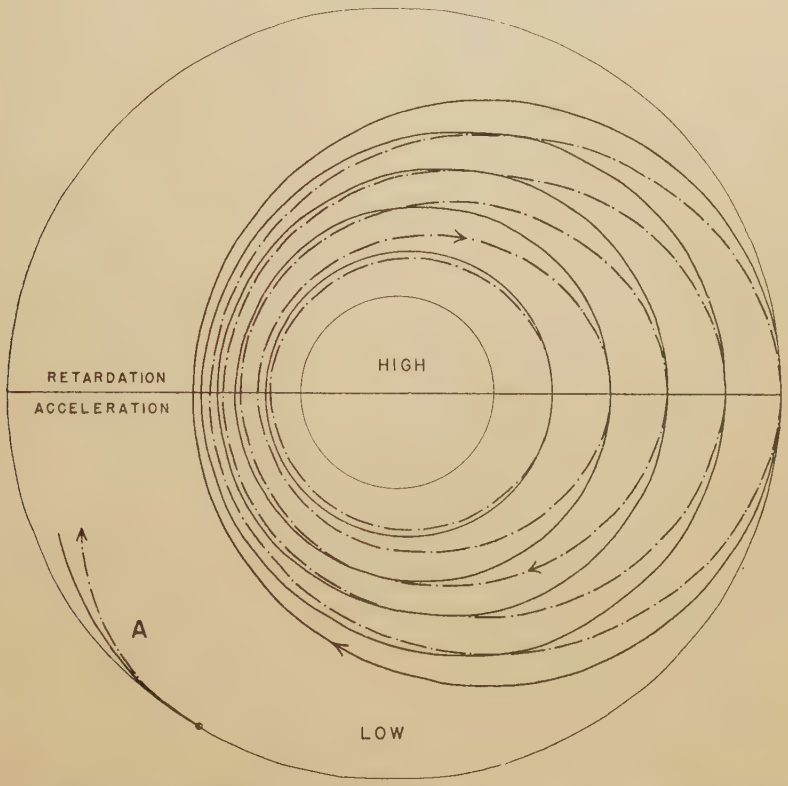

Figure 43. Qualitative illustration of probable relationship between dynamic contour lines and trajectories in an idealized oceanic circulation system on the northern hemisphere, subject to acceleration in low latitudes and retardation in high latitudes. Contour lines solid, trajectories dot and dash.

obtained by the dynamic method in so far as one is only concerned with approximate velocities and approximate directions of flow at any particular point, while a very great cumulative error would be likely to develop if the method were used for the determination of current trajectories over longer distances.

In a separate report the writer (Parr 1937c) has endeavored to show how the Caribbean Current with a known direction of flow, apparently must

ascend a dynamic slope in its passage through the Carribbean and Cayman Seas, and has ascribed this ascent to the effects of acceleration from above, caused by the steady trade winds of this region, with the suggestion that the case is probably only an illustration of the general principle that acceleration from above must cause an ascent of calculated dynamic gradient, retardation inversely causing descent. When acceleration or retardation proceeds from below due to topographic constriction or expansion, the writer further suggested that the effects should be the opposite of those caused by accelerating or retarding influences from above; acceleration from below resulting in a descent of calculated dynamic elevation, retardation from below in an ascent. This case of the Caribbean Current the writer considers so well established both in qualitative principle and in the actual example, which, due to topographic limitations, is not open to question with regard to the general direction of flow, that no repetition or further elaboration is necessary on these pages.

If, on the other hand, Jacobsen's illustration of the dynamic topography of the North Atlantic is approximately correct, we may find further, and perhaps even better, corroborative evidence in the north equatorial current, which when Jacobsen's and Meyer's maps (Fig. 41 and 42, see contour lines for 45 dyn. cm. relative elevation entered in Fig. 42 according to Fig. 41) are compared would seem to traverse the contour lines more or less at right angles, ascending westward to greater calculated dynamic elevations under the acceleration received from the trade winds.

If we now finally turn to a comparison between the dynamic topography of the sea surface of the North Atlantic as presented by Jacobsen (1929, Fig. 41 in this article) and the empirical surface current map prepared by Meyer (1928, Fig. 42 in this article), we get a convincing impression of the extent of the disagreement between dynamic contour lines and actual current trajectories, seeing that these run almost as frequently at right angles to each other as they run parallel.

In conclusion we might therefore attempt to deduct the general principle underlying this distorted relationship between contour lines and current trajectories. Under real conditions in nature, it is evident that both frictional retardation and acceleration by winds and other factors must influence the actual velocities of flow. In a stable circulation the sum totals of these retarding and accelerating factors over the entire circuit must exactly balance each other. If this was not the case, but total frictional retardation through the whole system exceeded the acceleration by wind, etc., the entire circulation would be running down to a stop. Or if the sum total of accelerating forces exceeded the retarding forces the speed of rotation would be constantly increasing.

If therefore we take an exact balance between total acceleration and retardation in a steady circulation for granted, we are thereby also forced

to accept as an unavoidable conclusion that acceleration and retardation will *not* balance each other over the greatest parts of the rotation system, since we *a priori* know that e. g. the accelerating effect of the steady trade winds must far exceed the accelerating effect of the weaker and more variable winds in other regions, and must even compensate for unfavorable winds over portions of the circulation system. It therefore seems evident that excessive acceleration in regions of strong favorable winds must cause an increase in kinetic or potential energy, to be dissipated again by excessive retardation in regions of less favorable wind conditions, and thus maintain the steady balance of the entire circulation system. It also seems fairly clear that the trade-wind belt must be the region of greatest excess acceleration and the northeastern quadrant (in the North Atlantic, at least) will generally be the region of greatest acceleration deficiency (see also Montgomery, 1936). Simplifying still further, one might say that the equatorial half of the rotation system will tend to be accelerated, the polar half retarded.

Let us now assume that we start with an ideal system of rotary flow, with equally spaced concentric dynamic contour lines also representing actual trajectories. Let the southern half (on the Northern Hemisphere) be subject to acceleration in excess of retardation. According to our previous conclusions (Parr 1937c, see p. 130 above) we would then expect the actual trajectories to begin an ascent of the original dynamic gradient. But by this ascent the current will also bring in towards higher dynamic elevations some of the water masses whose original distribution determined the concentric configuration of the dynamic contour lines. In other words, the originally concentric contours will also, under the influence of acceleration, have been drawn in towards the center of rotation of the entire system. Since the premise that the accelerated current must ascend, however, still holds, the deflection of the dynamic contour lines towards the center of rotation must be less than the deflection of the trajectories in a stable relationship between acceleration and dynamic topography, as shown at the lower left (A) in Fig. 43. It will hardly be necessary to elaborate the inverse reasoning for the descent of trajectories, and outward deflection of contour lines in the retarded half of the system.

If we now trace trajectories and deflected contour lines from common points of origin along the boundary between acceleration and retardation in the right half of the system, we get the picture shown in Fig. 43, in which the dynamic contour lines may form an excentric system of circles or of relatively broad oval forms, and the trajectories a similar system of narrower oblong curves intercepting the contour lines towards the center of circulation under acceleration and away from the center by net retardation.

Comparing Fig. 41 and Fig. 42, we find exactly the same general relationship between the dynamic contour line system of the North Atlantic, cal-

culated from actual observations, and the surface trajectories determined from empirical data, which are further overwhelmingly confirmed by the distributions of surface salinities and temperatures (allowing for cooling and mixing) (see Schott 1926, Pl. VIII and X) and by biological evidence from the distribution of eel larvae (Schmidt 1933, see Parr 1936a). When we also take topographic influences and the influence of cooling, etc. (Parr 1936a, see p. 126) into account, the agreement in general principle between the abstract Fig. 43 and the empirical Figs. 41 and 42 is so good that it seems quite warranted to offer this qualitative hypothesis of the effects of differentially distributed acceleration and retardation as a tentative basic explanation for the general relationship and distortion between dynamic contour lines and actual trajectories of flow in the major oceanic circulation systems.

LITERATURE REFERRED TO

See next article in this issue, pages 152–154.

ISOPYCNIC ANALYSIS OF CURRENT FLOW BY MEANS OF IDENTIFYING PROPERTIES

By

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With regard to the identifying properties by which, in the writer's opinion, the actual trajectories of ocean currents can be more confidently traced than by the dynamic topographic method (see preceding acticle, Parr 1938, a) nothing much needs to be said in the present article. To the extent that these properties *do* serve to identify water masses of the same derivation in the course of their movement and their gradual changes, and *do* differentiate these water masses from those of other origins, the validity of the conclusions to be drawn from a study of the distribution of these properties would seem beyond dispute. Investigations by the method of identifying properties further has two great and obvious advantages over the dynamic method of approach to the problem of actual trajectories, namely, 1) that they do not require the arbitrary assumption of a stationary datum level, and 2) that they are not confined to a consideration of components in two dimensions only, but deal with the ocean on a three-dimensional basis in accordance with actual nature.

The identifying properties most commonly used are of course salinities and temperatures, which may be studied separately, or usually to much greater advantage by the correlation method first suggested by Helland-Hansen (1918) and further elaborated and applied by Jacobsen (1929) and many others. When comparisons are made between calculated dynamic topographies and horizontal distributions of temperatures and salinities, however, one must not lose sight of the fact that one is actually comparing only two different elaborations of the same original data, temperatures and salinities forming the observational basis for the dynamic calculations. In consequence only a *disagreement* between dynamic topographies and the distributions of these two properties will have any conclusive bearing upon the corrections to be applied to the dynamic interpretation, while an agreement obtained from this comparison is entirely inconclusive unless the layer studied by its identifying salinities and temperatures has itself been excluded from the computation of the confirming dynamic topography (Parr 1937a, p. 56). Identifying properties not affecting the dynamic calculations therefore have several advantages over those considered in these computations. Unfortunately most of these independent properties are less conservative than the salinities and temperatures. Nevertheless, oxygen distributions in particular have already found useful applications in such

studies, and Seiwell's valuable report on the oxygen distribution in the western North Atlantic (Seiwell, 1934), in which reference is give to all earlier work, might perhaps be specially mentioned as an illustration of the excellent results obtainable from this source. Rossby's application of the correlation method to oxygen distributions as well as to temperatures and salinities (Rossby, 1936) is also of particular interest. It is obvious that the list of independently variable properties can be indefinitely extended as our knowledge of the chemistry of the sea advances.

Apart from the question of finding the most revealing identifying property, we also have the problem of selecting the form of presentation best designed to bring out the full hydrographic significance of the distribution of the properties which are known and available for discussion. The earliest method, still in use for its own particular purposes, was that of simply presenting the distributions in a series of horizontal levels of various depths. In such horizontal trans-sections, however, obliquity of the layers will introduce apparent discontinuities entirely obscuring the actual continuity of the layers themselves. Lately this method has to an increasing extent been replaced by a presentation of distributions in horizontal projection regardless of actual depth, which permits the actual continuity of a certain layer to be fully illustrated regardless of its vertical fluctuations (see particularly the clearly stated arguments for this method given by Wüst 1936). But even this procedure has its serious limitations in the difficulty of applying it to any other layers than those in which the property under consideration is consistently present either in a maximum or in a minimum concentration in all verticals.

With recognition of the fact that current flow, as distinct from mixing, must follow along isopycnal surfaces (Rossby 1936), however, it is clear that the attention should be directed towards a study of the distribution of identifying properties in surfaces of equal density, and in this form of presentation we will obviously be entirely free from the particular limitations applying to either one of the other two methods just described. Such isopycnic methods of investigation have already been developed and applied to meteorological problems by Rossby and collaborators (1937 b) under the designation of isentropic analysis. Here again it is evident that it would be of great advantage to be able to study the distribution of an identifying property which varies essentially independently of the density of the medium. In the atmosphere specific humidity has proved very valuable for such comparison, in the ocean the distribution of oxygen and similar properties (see above) not significantly affecting densities are naturally immediately thought of. But unfortunately no other property is so regularly subject of observation in oceanographic research as are the salinities and temperatures which both enter significantly into the determination of densities. In trying to apply isopycnic analysis to the ocean currents, it

therefore seemed particularly desirable to attempt to develop methods by which this analysis could be carried out on the basis of these two dependent variables alone. That the use of these dependent variables is not in itself objectionable if it will serve its purpose, is obvious from the great success and amazing precision of the temperature-salinity correlation method applied in a single dimension only (T-S correlation curves for single verticals of observation), since the study of the distribution of one or other of these variables along isopycnal surfaces is simply an extension of the T-S correlation method from one to two dimensions, the correlation with the other variable entering in the determination of the isopycnal surface. By integration of transverse profiles plotted on the basis of isopycnic distributions, the method can finally be extended to the third dimension as well, in the manner to be discussed in this article. From this reasoning it should be clear that the method of plotting salinities or temperatures against densities or isopycnal surfaces does not go beyond the premises of the simple method of T-S correlations alone, but merely extends the method to two or three dimensions, by plotting the variations of one variable against the correlation of both variables expressed in the densities which are the simplest and most obviously significant form of expression for this correlationship. It would of course also be very interesting to try other methods of expressing the variations of one against the correlation of both so as to obtain more than a one-dimensional picture, but this would of necessity involve the fitting of mathematical functions to the empirical forms of the T-S curves for single verticals, and a generalization of these functions for each type of water present in the region under investigation. An important contribution to oceanographic method pointing in this direction has been made by Iselin (1936) in his analysis of the distribution of deviations from the average T-S correlation curve for the North Atlantic. While this method has its own very valuable uses, it is desirable, in the isopycnic analysis of flow, to have the deviation of each single set of observations expressed not only by the separate deviation of each variable from its average for the observed separate value of the other, but also in terms of the relative deviation of each variable from its average for the observed correlation between both variables. This is what the plotting of either variable against the correlation of both, as expressed in the densities, will accomplish with a minimum of mathematical complication.

In the preceding we have repeatedly referred to isopycnic flow as an established fact, and before we go on the writer would like to point out that, apart from the surface layer in which densities may be changed by external influence (Parr 1936 a), the assumption of isopycnic flow is not a mere hypothesis or theory, but an axiomatic truth by the definitions of our hydrodynamic concepts, since, away from this surface layer, a liquid particle can only traverse an isopycnal surface, that is can only change its

density, by *mixing* (assuming diffusion to be negligible in comparison), while longitudinal *flow*, that is the essentially intact movement of a body of water as a stable unit with retention of its properties, on the other hand involves no change in density and must therefore *ipso facto* be isopycnic. Any component at right angles to the isopycnals is therefore by definition a mixing component as distinct from the trajectories of flow which must by the same definitions be isopycnal. This, however, is not to say that all isopycnic components are to be classified as components of flow, since the mixing processes will also have their greatest components along isopycnic surfaces (Parr, 1936b; Rossby 1936).

In this first attempt to investigate the possibilities of the methods herein suggested, it was found desirable to work with observations from a region in which the variation in the identifying properties of the adjacent water masses were greater than in the very homogeneous Caribbean Sea, in the study of which the writer is otherwise engaged. In the region of convergence between the Labrador Current and the North Atlantic drift off the southern end of Newfoundland bank the variations in identifying properties over short distances are at a maximum, and an excellent example for the study of the effectiveness of isopycnic analysis is here again available through the observations of the U. S. Ice Patrol.

DESCRIPTION OF PROCEDURE

The special set of observations to be discussed in the following were made at U. S. Ice Patrol stations 604–630, during June 19–29, 1926, and all the original data have been published by Smith (1927).

As a first step in the procedure salinities and temperatures were plotted against densities (σ_t)[1] for all stations (curves not shown). Salinities were then selected for further elaboration. To elaborate both properties from a plotting against density would obviously be meaningless in view of their mathematically rigid inverse correlationship at constant density values. Individual salinity profiles for selected isopycnals and for the free surface were then plotted for each section as shown by an example in Fig. 44. This plot is mainly done for the purpose of seeing whether the form of the curves will require the interpolation of maxima or minima beyond those occurring in the actual observations. From such profiles, and from the salinity-density curves, the horizontal distributions of salinity in various isopycnal sheets are then plotted as shown in Figs. 46–49.

In meteorological practice the distribution of identifying properties in isentropic surfaces is generally presented in a superimposition upon the topography of the isentropic surface with reference to elevation. In oceanographic usage the variations in the depth of the upper isopycnic surfaces

[1] See Montgomery 1937.

will generally tend to be too slight to be of any certain significance. The isopycnic isobaths have therefore only been drawn for the deeper surfaces of equal density in Figs. 47–48. In the case of the surface of $\sigma_t = 27.5$ the bathymetric chart is presented separately to avoid confusion (Fig. 50).

As a final step the minimum and maximum salinities occurring in each isopycnal surface within the entire region under investigation are then determined. We now assume that the maximum salinity in any isopycnal expresses the most unmixed state in which one of the original water masses contributing to the system occurs within the observed area, and that the minimum salinities expresses another original water mass in the same manner. Subtracting the minimum from the maximum, we then get the range of the difference between these two original components in the state in which

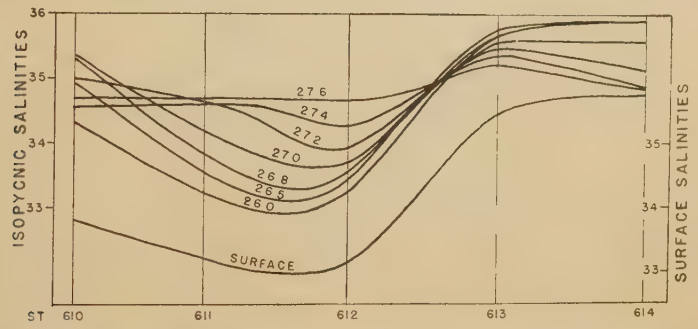

Figure 44. Salinity profiles for isopycnic layers of $\sigma_t = 26.0$; 26.5; 26.8; 27.0; 27.2; 27.4; 27.6 and for the free surface through Ice Patrol Stations 610–614. Salinity scale for isopycnic profiles at left, for free surface at right.

they enter our field of observation. Subtracting the minimum salinity in each isopycnal sheet from all values determined by observations and interpolations within the same density surface, and expressing these differences in per cent of the maximum difference within the isopycnal sheet, we should then obtain relative salinity values expressing the percentage presumably contributed by each original water mass to create the salinities actually observed in each surface of equal density, with 0 expressing one component in unmixed form and 100 expressing the other. The accompanying table gives relative salinity values in isopycnal surfaces for all stations considered in this example, and the station curves for relative salinities plotted against densities in situ (σ_t) are given in Fig. 51 A–E.

If we have only a single short series of relatively few stations, the usefulness of the method will be minimized by the fact that a number of observations corresponding to the number made at one station (although not necessarily all from the same station) will have to be given the value of 0,

Figure 45

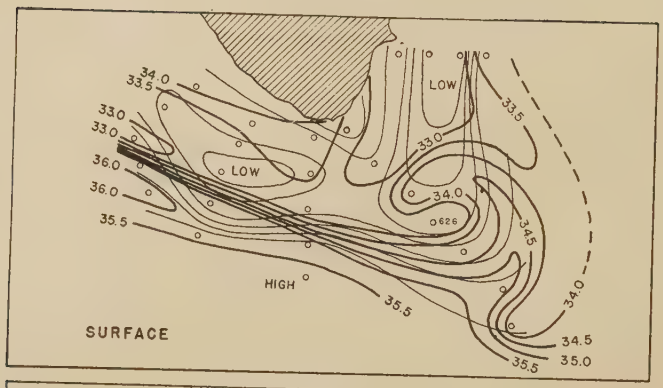

Figure 46

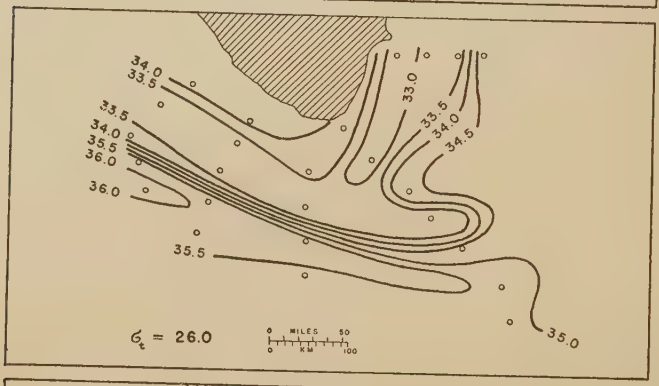

Figure 47

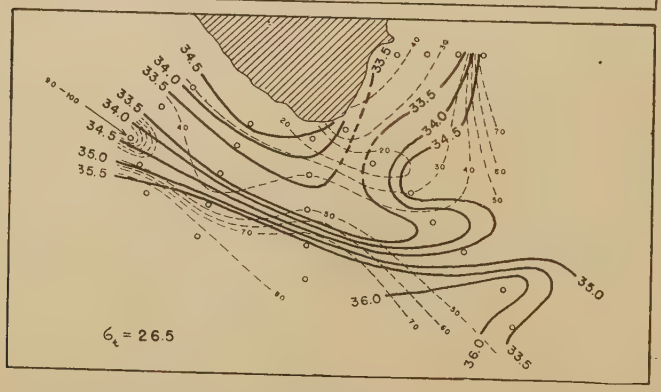

Figure 45. Surface salinities (heavy lines) superimposed upon the dynamic topography of the free surface with reference to 750 decibars (light lines) as drawn by Smith (1927, fig. 57) for Ice Patrol Stations 605–630.

Figure 46. Salinities in isopyncic surface of $\sigma_t = 26.0$. Isohalines for every .5°/₀₀.

Figure 47. Salinity distribution in isopycnic surface of $\sigma_t = 26.5$. Isohalines (heavy solid lines) for every .5°/₀₀. Isobaths (light broken lines) for every ten meters depth.

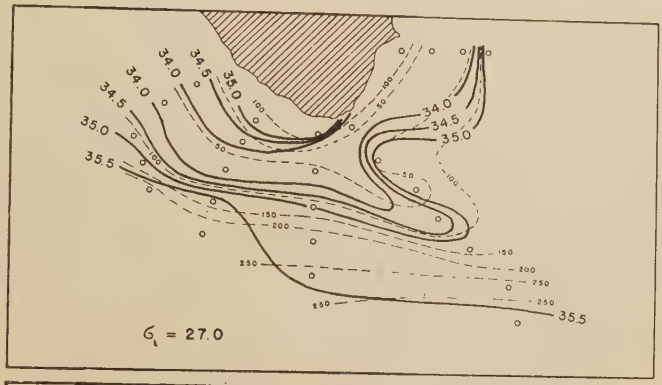

Figure 48

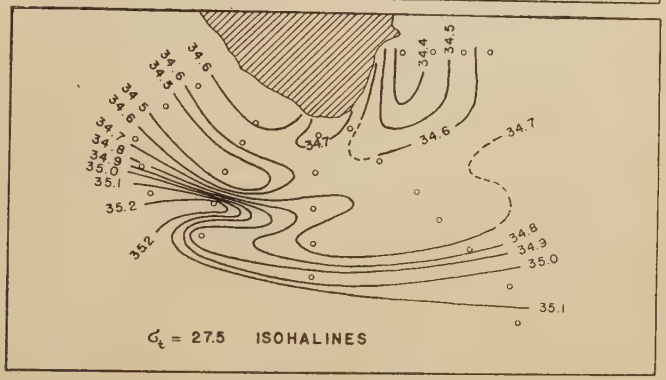

Figure 49

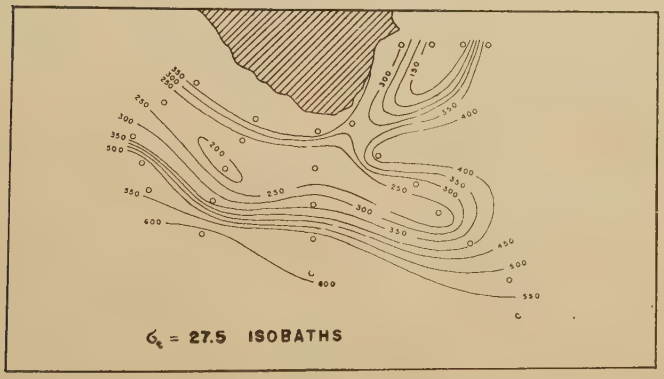

Figure 50

Figure 48. Salinity distribution in isopycnic surface of $\sigma_t = 27.0$. Isohalines for every .5°/oo. Isobaths for every 50 meters depth.

Figure 49. Salinities in isopycnic surface of $\sigma_t = 27.5$. Isohalines for every .1°/oo.

Figure 50. Bathymetric topography of isopycnic surface $\sigma_t = 27.5$. Isobaths for every 50 meters depth.

and an equal number the value of 100. That is, if a number of observations (r) is taken at each of a number of stations (n) out of the total number of observations (rn) only r (n-2) will actually give significant relative values,

RELATIVE SALINITIES IN ISOPYCNIC SURFACES
ICE PATROL STATIONS 604–630

Densities in situ	26.0	26.5	26.8	27.0	27.2	27.4	27.5	27.6
Maximum salinity	36.04	36.00	35.97	35.62	35.49	35.35	35.28	35.21
Minimum salinity	32.88	33.09	33.36	33.61	33.91	34.22	34.39	34.56
Salinity range	3.16	2.91	2.61	2.01	1.58	1.33	.89	.75
Station	*Relative salinities in per cent*							
605	100	97	92	100	100	100	85	56
606	88	56	87	92	80	49	46	35
607	24	39	48	64	66	34	30	24
608	9	9	7	6	7	9	12	19
609	40	34	27	22	19	17	26	21
610	45	63	74	86	69	26	21	19
611	11	15	19	27	46	29	25	17
612	11	11	7	4	1	4	9	13
613	88	92	88	98	98	85	100	100
614	93	96	98	98	77	50	52	40
615	75	94	93	91	80	60	66	45
616	97	93	91	85	61	36	31	17
617	19	3	10	70	64	41	38	24
618	25	22	19	18	19	23	29	36
619	35	41	58	76	59	40	40	32
620	24	17	7	3	10	17	25	32
621	2	20	35	83	50	26	26	25
626	39	55	66	82	58	35	30	43
627	11	18	22	28	35	26	27	21
628	60	52	49	85	75	41	37	29
629	81	100	100	84	75	61	72	60
630	68	83	89	94	82	70	82	71
622	0	0	0	0	0	0	0	0
623	6	7	5	4	6	5	7	15
624	9	11	9	8	8	14	21	14
625	52	61	65	72	56	16	34	21
604	7	13						

and even their relative values will have less significance in proportion to our lack of assurance about the significance of our limiting values with too few observations available. When the relative salinity-density curve for any particular station assumes the value of 100 or of 0 over any great density range, this does not mean that the waters at such a station are entirely homogeneous with reference to primary origin of one water mass

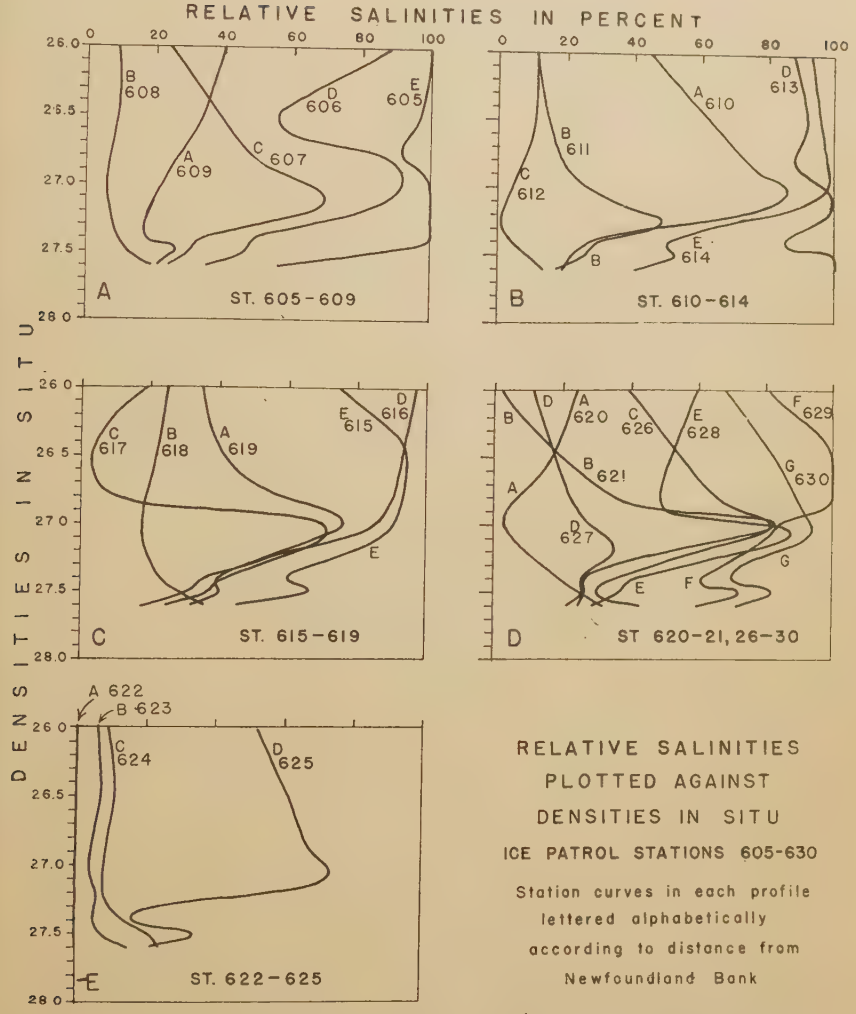

Figure 51. Relative salinities plotted against densities (σ_t) for separate profiles. Ice Patrol Stations 605–630. (Sequence of profiles from West [A] to East and North [E].)

or the other, but merely that over this density range such stations show the relatively least mixed state of the original components *as they arrive within the region of observation*, along isopycnal surfaces, although with reference to primary origin beyond this region it is most likely that these stations also would show variable degrees of already accomplished mixing over the entire density range.

From the relative salinity-density curves (Fig. 51 A–E) the relative salinity profiles can be drawn as shown in Figs. 52–56, in which relative

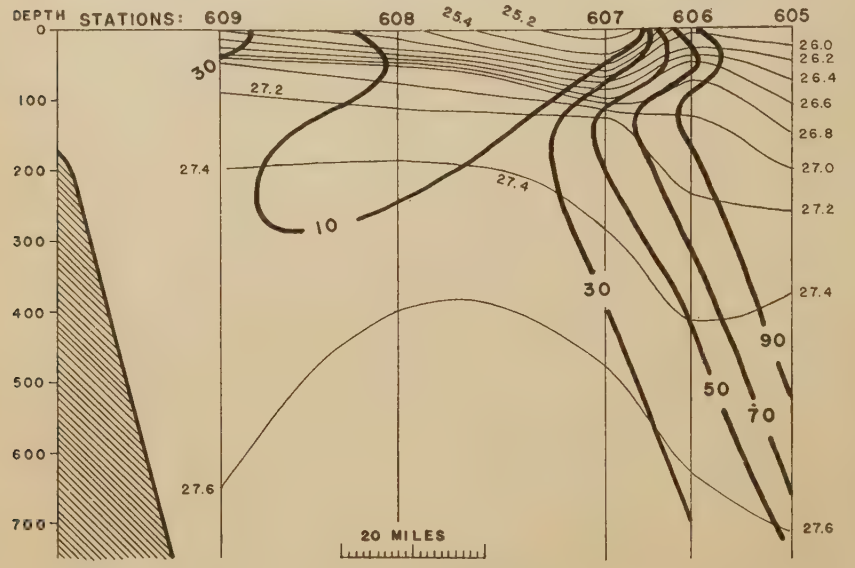

Figure 52. Profile of relative salinity curves (heavy lines) superimposed upon equal density curves (light lines). Ice Patrol Stations 605–609.

salinity curves are superimposed upon the isopycnals (figs. 52–53, 55–56) and upon the absolute salinity curves (fig. 54).

By integrating our relative salinity profiles (Figs. 52–56) we finally obtain a three-dimensional representation such as that roughly indicated in Fig. 57, in which the overlap between the heavily drawn 50% relative salinity contour lines for the Labrador Current (solid line, dotted area) and Atlantic waters (broken line, hatching) in the south and east, and between the Labrador Current waters and the waters off the slope of the bank (broken line, hatching) in the northwest, illustrates the actual overlap in the three-dimensional distributions of these water masses indicated by the relative salinity profiles.

DISCUSSION OF RESULTS

Figs. 46–49 will probably be sufficient illustration in themselves of the merits of the isopycnic method of analysis for tracing the actual distribution of water masses resulting from the movements of the ocean currents, and thereby also for outlining the general features of the current system itself.

It will be noted at once that the isopycnic salinity distributions for all surfaces of equal density here illustrated show a perfect agreement in indicating a major eddy formation in the eastern angle of divergence following the convergence between the Labrador Current and the North Atlantic drift. Both the position, and the indicated counter-clockwise direction of rotation of this eddy are in perfect accordance with every logical and empirical expectation in association with a convergence of this type. Nevertheless, no such eddy formation was indicated in the original dynamic topographic map of this hydrographic situation. While it may be true that, with this other evidence of the existence of such an eddy available, the dynamic contour lines might, without violation of the data, be redrawn so as to give a slightly better agreement with distribution of the water masses on certain points, such agreement would be *utterly unobtainable* for the western half of the picture, and the discrepancy between the identities of the water masses located in the incoming and in the outgoing contour lines in the North (Parr 1938a, this issue, p. 122) is entirely ineradicable. Since, furthermore, the dynamic map originally published (Smith, 1927, fig. 57, p. 115) is in every way a proper expression of customary methods of dynamic interpretation of current trajectories, it provides a perfectly fair basis for a demonstration of the limitations of this procedure in comparison with the results of the isopycnic analysis of flow (see Fig. 45).

The instructiveness of the relative salinity profiles (Figs. 52–56) with regard to the cross-sections of the various currents passing through the profiles in different directions is self-evident. In Fig. 54 the relative salinity distributions in two of the profiles are superimposed (in heavy lines) upon the pattern of the absolute salinity distributions (light lines). The much greater sharpness of definition obtained from the relative as compared with the absolute salinity profiles is obvious in these illustrations. One will particularly notice a complete lack of definition of the northern (left) boundary of the Labrador Current in the absolute salinity pattern for the upper 100–200 meters in Fig. 54, and the equally complete lack of definition of the bottom water (or mixing zone) rising to high levels at station 612. That the absolute salinity distributions in other parts of these particular profiles might in themselves alone lead to approximately the same (but less precise) interpretations of current boundaries, as do the relative salinities, is due to the fact that the differences in the identifying properties of the separate bodies of water in this particular instance are so great that inversions occur both in the horizontal and in the vertical salinity gradients. It

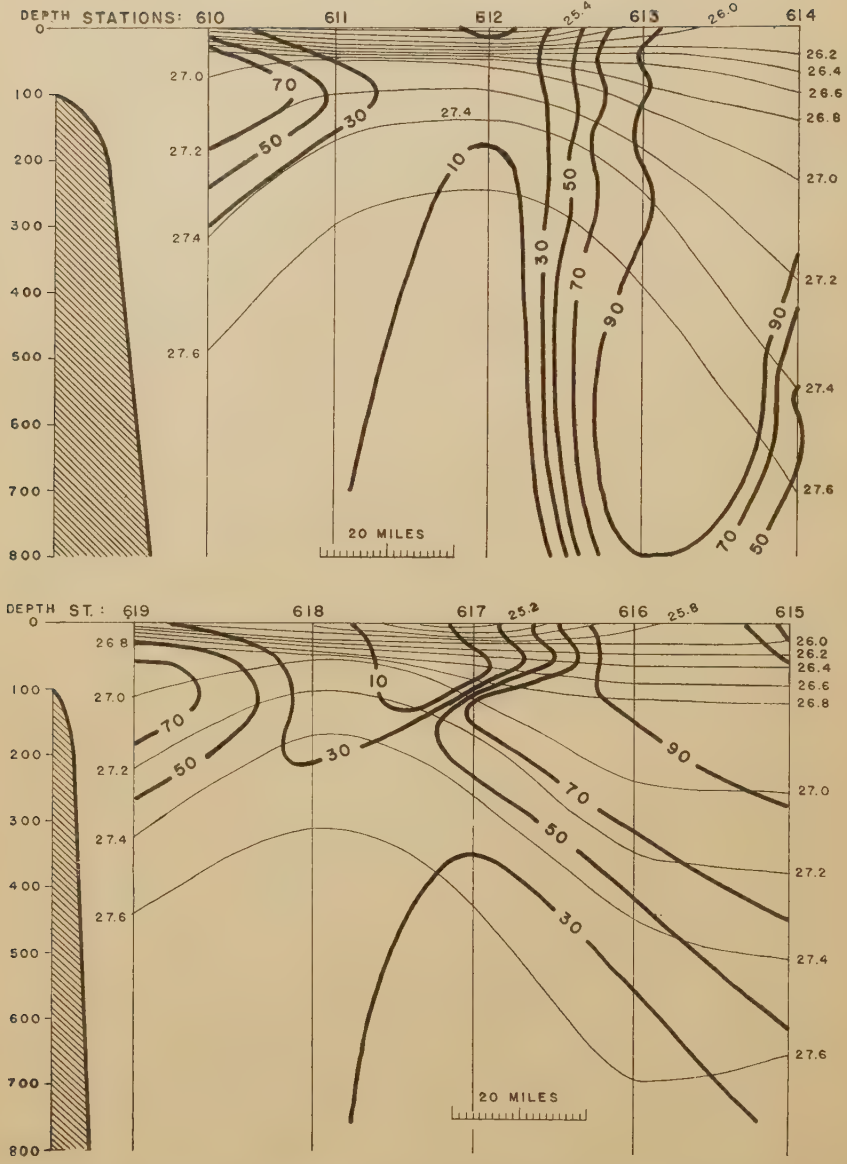

Figure 53. Profile of relative salinity curves (heavy lines) superimposed upon equal density curves (light lines). Ice Patrol Stations 610–614 (above) and 615–619 (below).

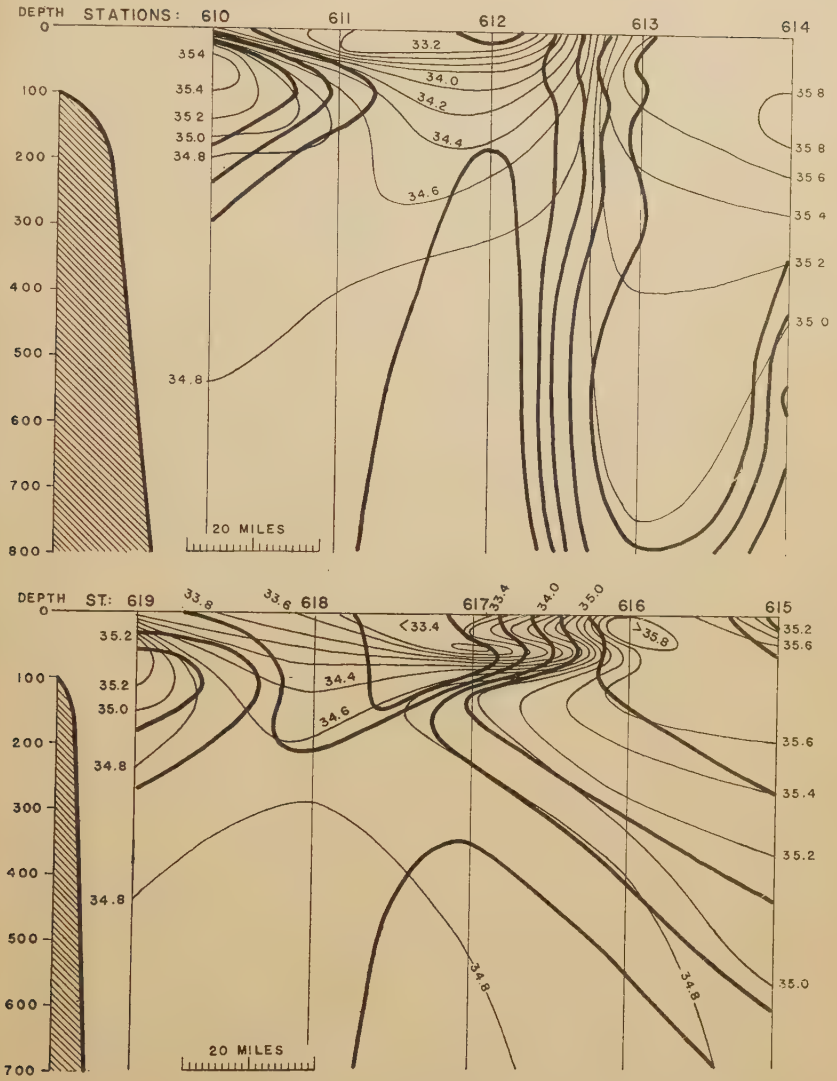

Figure 54. Profile of relative salinity curves (heavy lines) superimposed upon absolute salinity curves (light lines). Ice Patrol Stations 610–614 (above) and 615–619 (below).

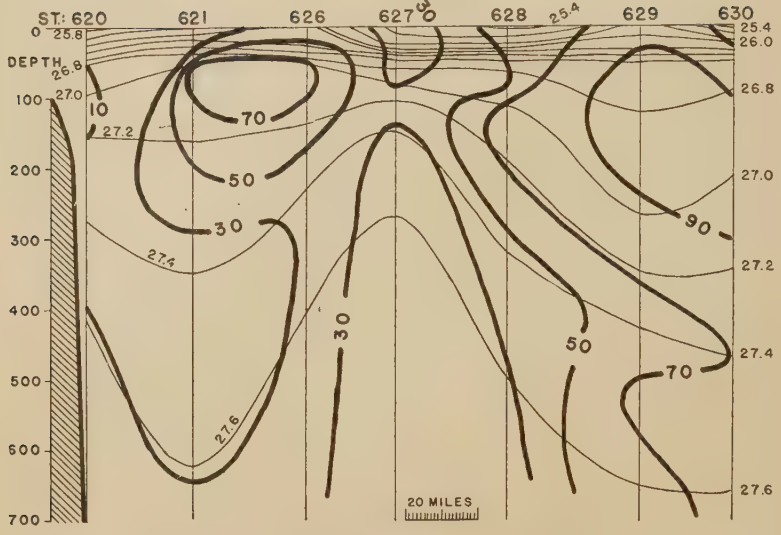

Figure 55. Profile of relative salinity curves (heavy lines) superimposed upon equal density curves (light lines). Ice Patrol Stations 620–21; 626–630.

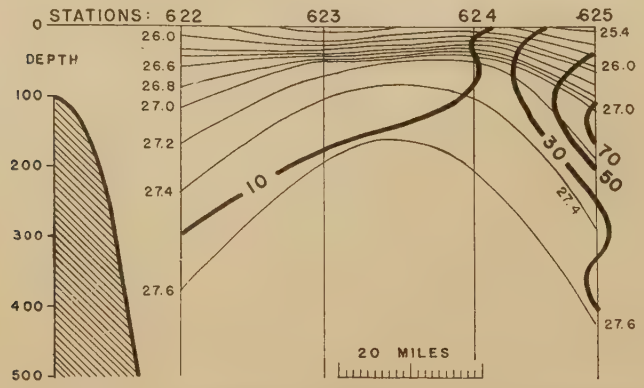

Figure 56. Profile of relative salinity curves (heavy lines) superimposed upon equal density curves (light lines). Ice Patrol Stations 622–625.

is clear, however, that inversions in the lateral (as distinct from the horizontal) salinity gradients along the undulating isopycnal surfaces may perfectly well occur without leading to corresponding inversions either in the horizontal or in the vertical gradient. Under these conditions, which are likely to develop to a greater extent the less sharply distinct the various bodies of water are, the definition of the current boundaries will tend to disappear entirely from the absolute salinity profile, as they have disappeared for this

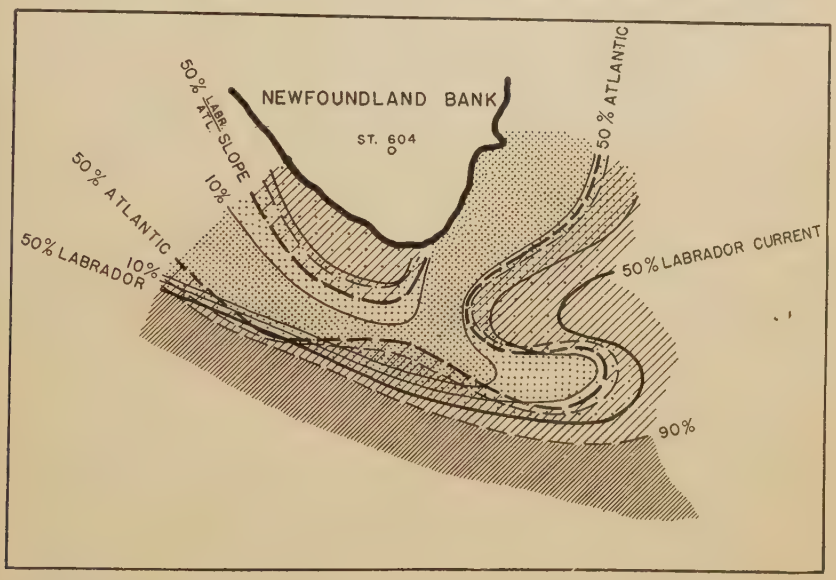

Figure 57. Distribution of relative salinities (10; 30; 50; 70 and 90%) regardless of depth. Dotted area and solid contour lines for relative salinities of less than 50% (Labrador Current). Hatched area and broken contour lines for relative salinities of more than 50% (Atlantic water). 50 per cent limiting boundaries of each system heavier than other contours.

very reason in the particular portions of the two profiles here discussed (see above), while the definitions of the boundaries in the relative salinity profiles remain equally sharp under all circumstances. This variation in the relationship between the relative and the absolute salinity distributions may be generalized in the statement that so long as the absolute salinities as a whole, that is both minimum and maximum salinities, vary with density *in situ*, relative and absolute isohalines can only run parallel with each other when the absolute isohalines are parallel with the isopycnals, but must have a component at right angles to each other whenever the absolute isohalines have components at right angles to the lines of equal density. Since absolute salinities almost always vary with density (similarly as with

depth) and since the isopycnal surfaces will only exceptionally also be surfaces of equal salinities for any great distance, it follows that relative and absolute salinity profiles will only in extremely rare and limited cases, if ever, present identical pictures. It is therefore evident that the relative salinity profiles or any other form of presentation of relative salinities over a range of variations in density (e. g., in charts drawn regardless of depth and density such as Fig. 57) will always have a superior significance of their own, with a more direct bearing upon the problems of flow than that inherent in the absolute salinity distributions.

If, on the other hand, a presentation of salinities with reference to constant density is desired, it is obvious that any distinction between relative and absolute salinities would be purely numerical and entirely pointless. For comparison of relative salinity distributions from one isopycnic chart to another all that is required will be a rectilinear conversion curve for each surface from minimum salinity, as 0%, to maximum salinity, as 100%, but the isopycnic pattern of relative isohalines will be identical with the isopycnic patterns of the absolute salinities.

Concerning the relationship between bathymetric topography and the distribution of identifying properties in an isopycnic surface, it is the general expectation in meteorology to find a parallelism or at least a fairly distinct similarity between the pattern of the topographic contour lines and of the lines of equal concentration of the property considered. In our example of the hydrographic application of isopycnic methods of analysis, the surface of $\sigma_t = 26.5$ is apparently still not far enough below the free surface of the sea to escape the distorting influence of the variable forces such as wind, etc. acting from above upon the bathymetric characteristics of the surface layer itself. In the isopycnic surface of $\sigma_t = 27$ (Fig. 48), however, we find a fair, but far from perfect, agreement between bathymetric topography and identifying properties, in accordance with the just mentioned expectations. This agreement is also repeated in the general pattern of the still deeper isopycnic surface of $\sigma_t = 27.5$ (Figs. 49 and 50), with a notable exception, however, in the southwest portion, for which the writer has no explanation to offer at the present time.

With regard to mixing, the superimposition of the relative salinity profiles upon the density profiles as in Figs. 52–56 gives us at least a valuable qualitative indication of the distribution of mixing processes in terms of the relative significance of their horizontal and vertical components. Thus the manner in which the relative salinity curves intercept the density curves in the deeper layers at the right margin of every profile is probably indicative of a considerable vertical exchange in and below the deeper portion of the North Atlantic drift. When the relative salinity curves follow an almost isopycnic course as they do over considerable distances from the center of the profiles to the lower right in all sections, and also towards the lower

left in the more easterly sections (Figs. 53–54) it is probable that both lateral and vertical mixing must be relatively very slight in proportion to longitudinal isopycnic mass transport. When finally the intercept of the relative salinity pattern with the density pattern takes the general form characteristic of the upper layers throughout, and of the central column to considerable depths in each profile, a predominance of isopycnic mixing is obviously indicated, with its magnitude at least qualitatively suggested by the spacing of the relative salinity intercepts along the isopycnals.

The isopycnic analysis of the distribution of identifying properties may also be of special usefulness in locating centers of vertical motion. As already brought out in meteorology by Rossby and others, the isolated occurrence of high or low concentrations, that is, of areas surrounded by closed curves of equal concentration of the identifying property, in any isopycnic surface is *ipso facto* evidence of vertical motion. When variations in the identifying property studied do not significantly affect the densities, the case is comparatively simple, but when for such reasons as those advanced in the preceding (p. 135) we want to base our analysis upon dependent variables, certain complications arise. In either case it is necessary to investigate the circumstances under which the presence of such closed curves in an isopycnic surface may be due simply to vertical or lateral mixing, and when it may be considered to provide unequivocal evidence of vertical convections passing through the surface in question.

Unfortunately our example from the convergence of the Labrador Current with the North Atlantic drift does not provide an illustration of such closed curves, but the principles involved can be easily demonstrated on an abstract case. If we have a temperature-salinity correlation curve such as ABC in Fig. 58, it is evident that vertical mixing confined to one of the straight regions of this curve, between A and B or between B and C, cannot produce any changes either in the salinity or in the temperature of any isopycnic surface. If, on the other hand, vertical mixing occurred around the inflection point B, the T–S correlation curve might change from ABC to AB_1C and both temperatures and salinities would increase in the isopycnic surface of $\sigma_t = 28$. In the isopycnic surfaces around any curved portion of the correlation curve between two identifying properties of which at least one varies with density,* areas isolated upstream with reference to the distribution of at least one of these properties may develop simply through vertical mixing. If the vertical mixing is localized, it is further evident that these areas may also become isolated downstream by subsequent lateral mixing, and may thus be circumscribed by completely closed curves without giving evidence of anything but vertical and lateral mixing. To give conclusive evidence of actual convection through an isopycnic

* Whether by direct causal relationship or indirectly is entirely immaterial in this connection.

surface, as distinct from mixing, the occurrence of isolated areas of maximum or minimum abundance of properties which vary with density in any significant manner whatsoever must be associated with a complete separation between the correlation curve of two such properties defining their relationship within the isolated area and that characteristic of their relationship in the surrounding waters, or by a *reduction* in the radius of curvature* of the correlation curve if the curve within the isolated area has an intercept with the correlation curve of the surrounding waters both above and

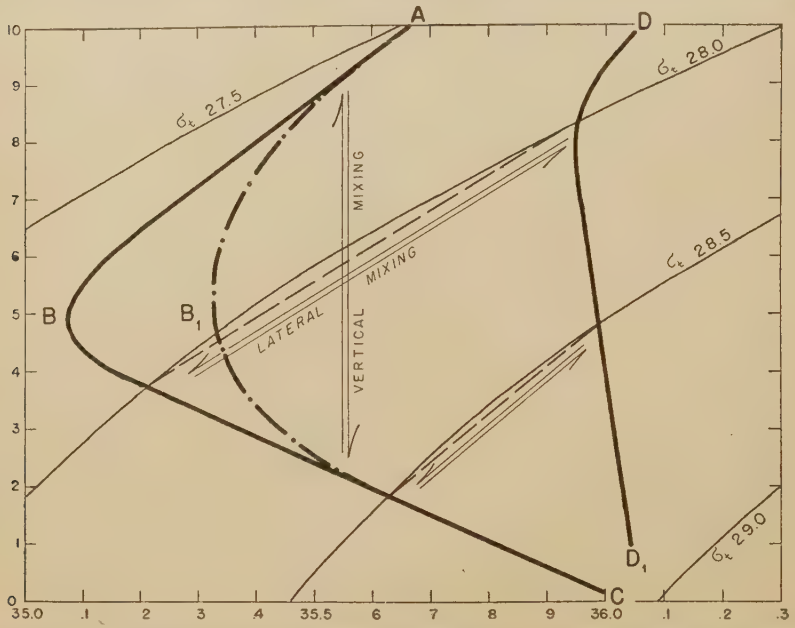

Figure 58. Effects of lateral and vertical mixing upon the isopycnic distribution of salinities and temperatures. See text.

below the isopycnic surface in which the isolated maximum or minimum appears.

With reference to the effect of lateral mixing upon isopycnic distributions, it is of interest to note that if we base our analysis upon temperatures and salinities, and lateral mixing occurs along isopycnic surfaces between two bodies of water characterized by separate T–S curves, such as for instance *ABC* and DD_1 in Fig. 58, the *rectilinear* gradient along the actual mixing trajectory must result in a *curved* gradient along the isopycnals for reasons which will be abundantly obvious in the illustration. That is to say, lateral

* The radius of curvature of a straight line being infinite.

mixing cannot actually be strictly isopycnic for the reason that the mixing products between *different* waters of equal density do not themselves remain of constant density (see also Montgomery 1937). If a sufficient number of sufficiently accurate determinations could be made, it seems indicated that the form of curvature of the isopycnic gradient between adjacent

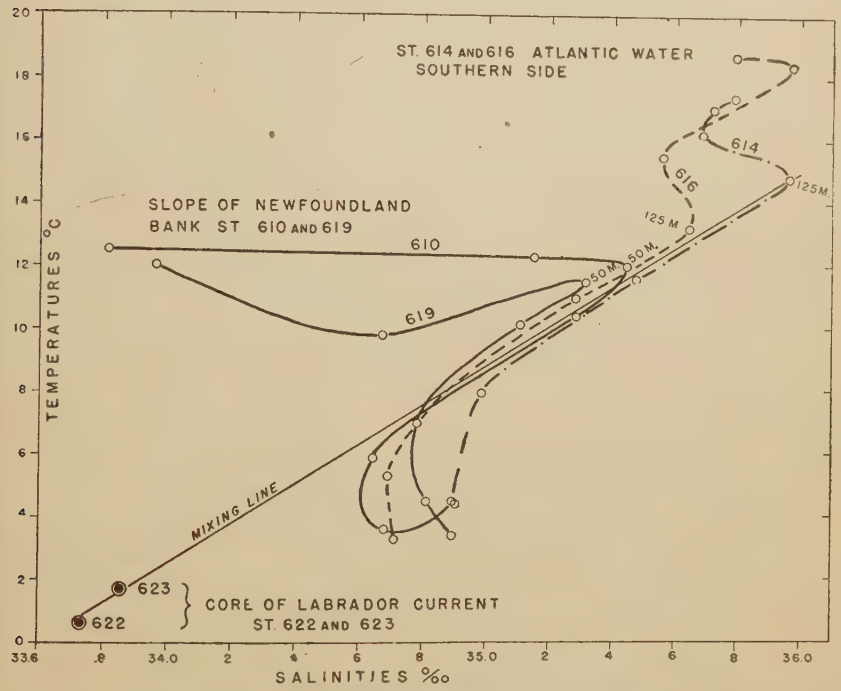

Figure 59. Temperature-salinity correlation curves for Ice Patrol Stations 614 and 616, representing Atlantic water; Stations 610 and 610, representing water off the slope of the bank; and correlation points for the minimum temperature layer in the Labrador Current (Stations 622 and 623).

bodies of water might provide a valuable clue to the lateral mixing processes themselves, but such accuracy is probably not yet obtainable in practice.

Concerning the apparently separate body of water of higher relative salinities occurring immediately off the slope of the Newfoundland bank, it is of interest to notice from Fig. 59 that its temperature-salinity correlations agree very closely with those obtaining in the North Atlantic drift, and that the relevant portions of the T–S curves, both for the North Atlantic drift in the south, and for the slope water in the north are pointing with an amazingly perfect fit in a straight line towards the correlation values obtained for the cold core (minimum temperatures) of the Labrador Current in the most northern profile.

Several possibilities thus suggest themselves for the derivation of this intermediate slope water. It may be the disintegrating end point of a slowly moving northern branch of the Atlantic drift, originating through a bifurcation farther to the eastward around a final submergence of the branch of the Labrador Current which still remains in the upper layers in the region here investigated. Or the slope water may be an accumulation from occasional eddy transfer from deeper levels in the Atlantic drift to higher levels off the bank, taking place *below* the Labrador Current, which would not require the assumption of any changes by mixing after the slope water has arrived in its northern location. Or it may finally and more probably arise from occasional eddy transfer (or regular eddy transfer in locations which have escaped observation) *over the top* of the Labrador Current, facilitated by the extremely sharp pycnoclines in its upper central portion. This would bring towards the slope of the bank water of higher salinities than those actually observed in the northern slope water (see Figs. 52–56), but the observed values would quickly develop through mixing *in situ* and in the course of the eddy transfer.

SUMMARY

An isopycnic method of analyzing current patterns by means of relative salinity distributions is suggested. As an example it is shown how this method serves to demonstrate the existence southeast of the Newfoundland bank of a large and deep eddy movement, the presence of which is entirely undisclosed in the dynamic topography of the free surface calculated with reference to a horizontal datum level.

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ON THE PROCESS OF UPWELLING

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The phenomenon of upwelling which is present along many coasts has been discussed from various points of view. A general explanation of the phenomenon has been given by application of Ekman's theory of wind-driven currents (Thorade, 1909; McEwen, 1912), and emphasis has been put on the velocity of the vertical motion (McEwen, 1934) or on the depth from which water is brought to the surface (Gunther, 1936; Sverdrup, 1930). Defant (1936) has discussed the horizontal and vertical motion and the distribution of density within the region of upwelling off the coast of southwest Africa, but his analysis is based on average surface conditions and on the results of a few, widely-spaced oceanographic stations. His schematic representation of the currents in a vertical section at right angles to the coast appears misleading, since the vertical motion shown in the region of upwelling necessitates the existence of a horizontal divergence within this region. But a divergence is indicated only at the offshore border of this area.

A more detailed analysis of the process of upwelling off the coast of California can be undertaken by means of observations which were made by R. H. Fleming and E. G. Moberg of the Scripps Institution of Oceanography on board the "Bluefin," the vessel of the California State Fish and Game Commission during three cruises in March, May, and June, 1937. On all cruises a section was located nearly at right angles to the coast of California off Port San Luis (Lat. 35° 10′ N., Long. 120° 45′ W. See figure 60). The stations of March 25–27 were reoccupied on May 5–6, and on June 26–27. Although the exact locations differed slightly, we shall in the following discussion deal with the data as if the locations were identical.

Figure 61 shows the dynamic profiles on cruises I and II of the 0-decibar surface, the 100-decibar, and the 200-decibar surfaces relative to the 500-decibar surface. It is seen that above the 200-decibar surface a very conspicuous change took place: on March 25–27 the surface rose more or less irregularly from the coast to a distance of 180 km. from the coast, but on May 5–6 the surface was nearly flat from the coast out to station 4, 100 km. from the coast, and then rose steeply, not less than 0.116 dyn. m. in the 28 km. between stations 4 and 5. The profile of the 100-decibar surface was

* Contributions from the Scripps Institution of Oceanography. New Series, No. 15.

changed similarly but less, whereas the 200-decibar surface remained nearly unaltered. It is evident that these changes must be due to a transport away from the coast and a banking-up offshore of the light surface water, and it is probable that the transport is caused by wind.

The wind direction and velocity in latitude 35° N, longitude 122° W, was derived from the weather maps constructed twice daily at the Lindbergh Aerological Station (San Diego). The direction of the wind was assumed to

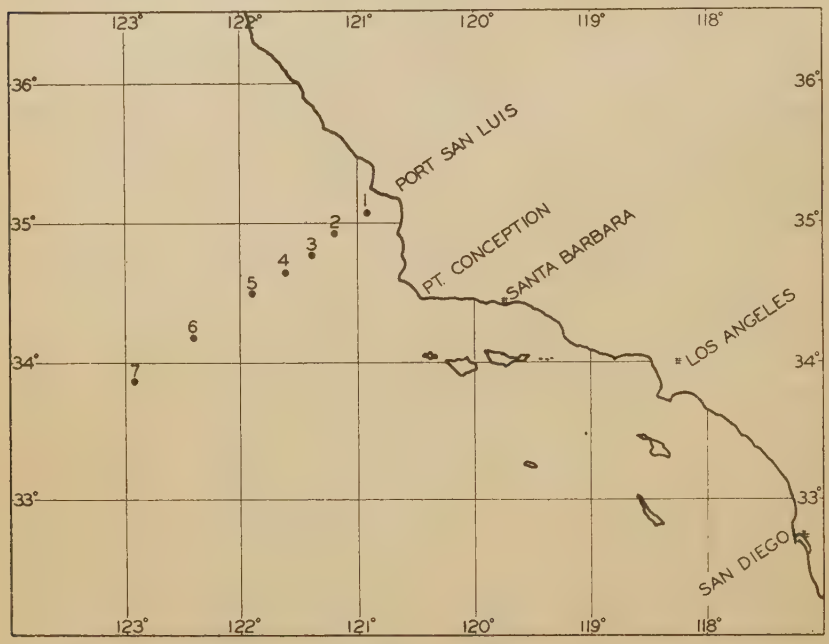

Figure 60. Chart showing the location of stations occupied on the northern line during the "Bluefin" cruises in March, May, and June, 1937.

deviate 17° from the direction of the isobars and the velocity was obtained by multiplying the gradient wind by the factor 0.62 (Rossby, 1936). During the five days prior to March 25, the average wind was 6.0 m/sec from S 85 W and, thus, the wind had blown from a direction which would transport the surface waters parallel to the coast. On March 23 and 24 the wind was strong enough to delay the beginning of the cruise. On March 25 the wind changed to NNW and during the following six weeks it blew from directions between N 55 W and N 20 E, occasionally reaching velocities as great as 12 to 14 m/sec. During the period of 41 days between March 25 and May 5 the average wind was 6.7 m/sec from N 23 W, and this wind which was nearly parallel to the coast would cause a transport of the surface layers

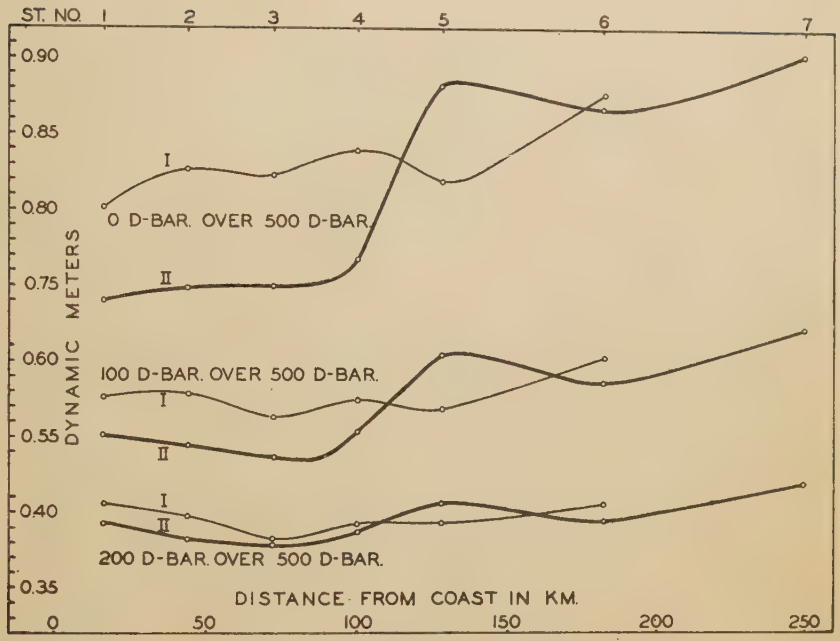

Figure 61. Profiles of isobaric surfaces relative to the 500-decibar surface on cruises I and II.

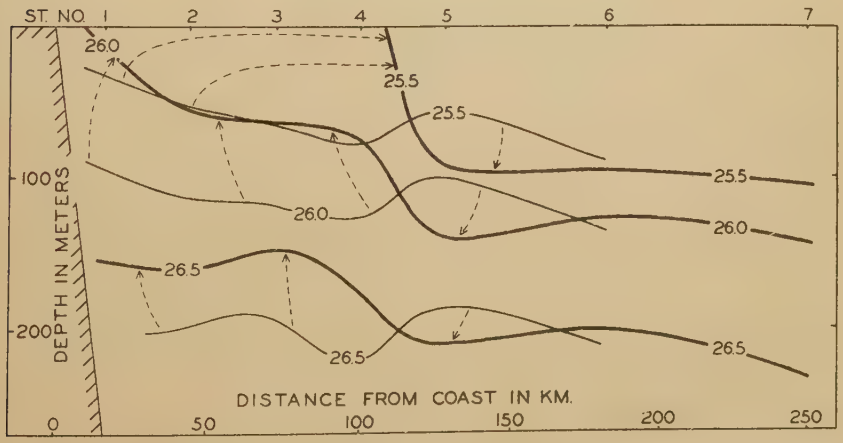

Figure 62. Vertical distribution of density (σ_t) on cruises I (thin lines) and II (heavy lines).

away from the coast. That such a transport took place is evident from a comparison of the distribution of density. The thin lines in figure 62 show the distribution of density on March 25–27 represented by means of σ_t which, in this case, give sufficient accuracy. The heavy lines show the corresponding distribution of May 5–6, and the arrows indicate the general character of the motion which may have brought about the new distribution. It is quite evident that the light surface-water has been transported outwards and banked up at a distance of about 100 km from the coast.

A somewhat different picture is obtained by a detailed analysis of the data. From the vertical displacement of the temperature, salinity, and

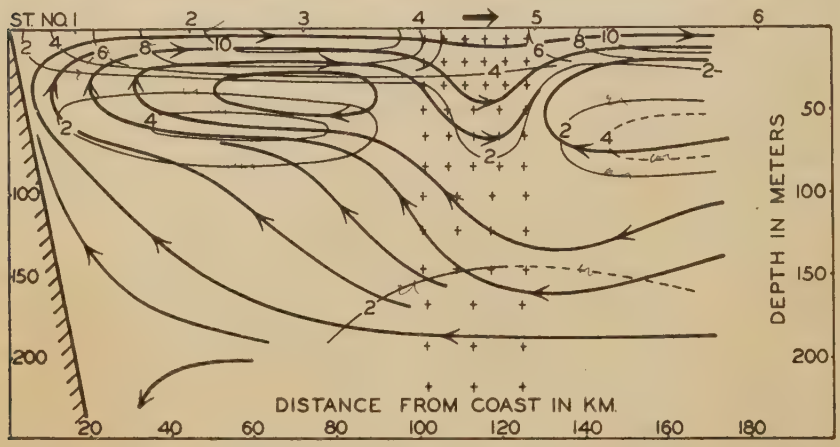

Figure 63. Computed vertical circulation in the time interval between cruises I and II. Direction shown by heavy lines with arrows; horizontal velocities shown by thin lines; areas with x's represent region with swift current parallel to the coast and away from the reader.

oxygen curves the average vertical velocities during the 41-day interval were computed, assuming that at the surface the average vertical velocity was zero. Knowing the vertical velocities, the horizontal flow was found by means of the equation of continuity, assuming uniform velocities parallel to the coast. The details of the computations will be reported elsewhere. Here only the final result is shown in figure 63. The lines with arrows represent the average stream lines in the 41-day time interval, and do not represent trajectories. The scale is arranged to show the inclination of the stream lines exaggerated in the same proportion as the vertical scale of the section (400 : 1).

The horizontal velocities are indicated by curves. The computed maximum offshore velocity is 11 cm/sec, in good agreement with the probable velocity of the wind current. The average outward velocity of the boundary

region between the upwelled water and the offshore surface water must, however, in the period March 25–27 to May 5–6, have been only about 3 cm/sec since water, which on March 25–27 was found near the coast, was on May 5–6 found at a distance of 90 km. from the coast. At a distance of 75 km. from the coast the outward transport above a depth of 40 meters, as computed from the equation of continuity, is 26×10^3 cm^3/sec, whereas a value of 25×10^3 cm^3/sec is obtained from the average tangential stress of the wind in the 41-day period, assuming a roughness parameter of 0.6 cm. (Rossby, 1936). The thickness of the upper homogeneous layer is 41 meters and Rossby's theoretical value of this thickness is 39 meters (Rossby and Montgomery, 1935). These agreements lend support to our representation of the flow.

The picture shows several striking features. Between the coast and the banked-up light surface-water we find a cellular circulation which reaches to a depth of about 80 meters and within which the water flows away from the coast above 40 meters, and towards the coast below 40 meters. Near the coast the motion is ascending, near the offshore boundary it is descending. The offshore boundary is, however, moving slowly away from the coast and the inner cell is, therefore, being fed from below by water which flows toward the coast. This inflow appears to take place at depths not exceeding 200 meters.

Figure 64 shows the profiles on cruises II and III of the 0-decibar, 100-decibar, and 200-decibar surfaces relative to the 500-decibar surface. It is evident that on cruise III the light offshore water was encountered at an even greater distance from the coast, but the banking-up effect does not appear so strikingly. When drawing the profile in the simplest manner, we find a gentle rise of the surface between stations 5 and 7, but between these stations we may actually have a steep rise on a short distance, as indicated by the dashed curves. Assuming that such was the case, we find that the boundary region has moved 70 km. away from the coast in the 51 days between the cruises, and that inside the boundary region an eddy of a diameter of about 75 km. is present.

A comparison of the density distribution (fig. 65) shows that near the coast some upwelling has continued and that vertical motion has taken place directly inside of the boundary, but between these regions of vertical motion water masses of a different character have been embedded. A closer study indicates that these water masses are similar to those present at the coastal side of the boundary region on cruise II or, in other words, that an eddy had detached itself from the coastal side of the boundary. The low density of the water directly inside the boundary on cruise III (between stations 5 and 6) is due to heating, since the salinity is high. It has not been attempted to calculate any vertical or horizontal velocities since the processes are too complicated.

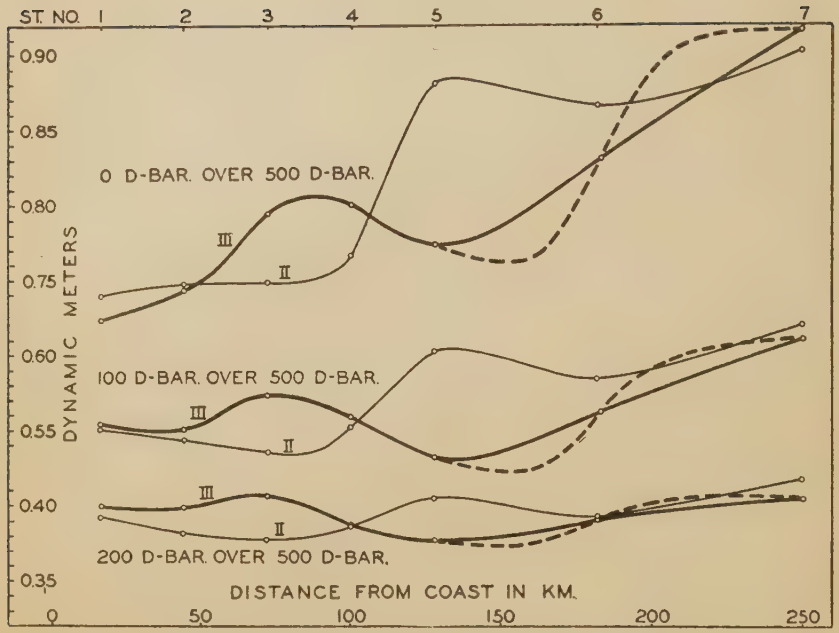

Figure 64. Profiles of isobaric surfaces relative to the 500-decibar surface on cruises II and III.

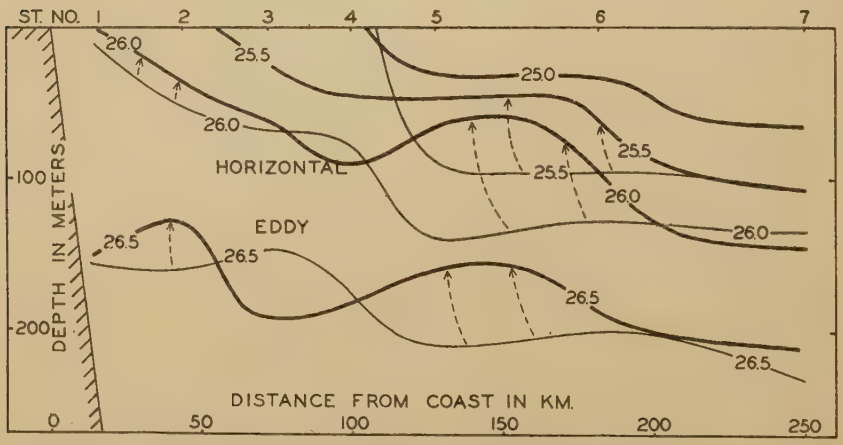

Figure 65. Vertical distribution of density (σ_t) on cruise II (thin lines) and III (heavy lines). Arrows indicate types of motion which have brought about the changes in the distribution.

In the period May 5–6 to June 26–27 the wind was weaker and from a more westerly direction. In figure 66 average 5-day values of the component of the wind parallel to the coast (from N 20 W) have been plotted for the time interval March 20 to June 28. It is seen that conditions were more favorable to upwelling between cruises I and II than between cruises II and III, and that between the latter cruises there was a period in which the component parallel to the coast was nearly zero. It is possible that the variable wind conditions in the time interval between cruises II and III were responsible for the break-down of the boundary region and the development of the eddy, but it will be pointed out that other possibilities exist.

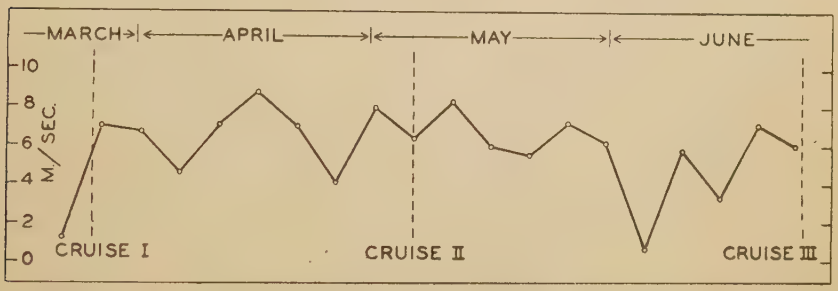

Figure 66. Component of wind parallel to the coast (from N 20° W in the period March–June).

It does not seem feasible to make the described conditions the subject of a mathematical analysis, but certain features can be dealt with by simple reasoning. Let us assume that a tangential stress directed parallel to the coast and away from the reader is exerted upon the surface of the water within which the distribution of density is that shown by the thin lines in figure 62. Under these conditions the uppermost light water will be transported away from the coast and must be replaced by water of greater density from some sub-surface depth. The upwelled water will in turn be carried away from the coast and, since this water has a greater density, a current parallel to the coast must develop and flow in the direction away from the reader. If within this current the velocity decreases with depth, as was the case on May 5–6 (compare the slopes of the 0-decibar surface and the 100-decibar surface and observe that the profiles of these surfaces cannot very well be drawn with the same slope), part of the tangential stress of the wind will be balanced by the stress due to the shearing motion of the current. This means that the area within which the current has been developed must move offshore at a speed which is smaller than the speed by which the surface layers inside the current are transported outwards since there the full amount of the stress of the wind will be active. Thus, the

region of the current will act as a barrier to the outward transport and inside this barrier a cellular circulation must develop. Since the barrier is moving, this cellular circulation must either become thinner or it must be fed from below. It is not probable that it will become thinner since the depth of the offshore flow must be dependent on the wind velocity and be relatively independent of the motion of the barrier. We conclude, therefore, that water from greater depth is being drawn into the inner cell as the barrier moves away from the coast.

Outside the region of the current we must obtain a somewhat similar system. We must find ascending motion near the barrier which will act as a secondary coast, but the ascending motion is restricted to the upper layer of light water and has not the character of upwelling, that is, no deeper water is brought to the surface. Below the top layer we must find descending motion feeding the current which flows towards the coast below the boundary layer. If this were not the case, it would mean that the effects of the upwelling could be detected at very great distances from the coast.

The region of the current parallel to the coast must be a region of intensive mixing since it moves away from the coast at a velocity which is smaller than the surface velocity inside the region, and since the kinematic boundary condition, that the velocity normal to the boundary shall at all levels be equal to the velocity of the boundary, cannot be fulfilled. This may explain why our data cannot very well be interpreted as showing a sloping boundary surface between the light and the heavy water such as should be expected if we were dealing simply with the boundary between two currents, but that the simplest interpretation shows a region of transition characterized by a strong current parallel to the coast within which the velocity rapidly decreases with depth. Observations within such a current are, however, desirable in order to establish this feature definitely.

It is perhaps conceivable that this decrease of velocity may become so great that the stress of the wind is completely balanced against the stress due to the shear in the water. In this case the region of the current would take the character of a solid stationary barrier and inside of the barrier a stationary cellular circulation could be present. It seems, however, more probable that a stationary stage is not reached and that conditions such as were found on cruise II cannot exist for any length of time. A change in the wind velocity or direction which leads to a change in the transport of the surface waters may alter conditions to such an extent that large eddies develop on the boundary between the upwelled water and the offshore surface water. In our case it must also be considered that the current flows from north to south in a region within which the deflecting force of the earth's rotation changes rapidly with latitude. This may lead to unstable conditions and give rise to the formation of eddies. It also should be considered that, since the kinematic boundary condition is not fulfilled, inten-

sive mixing must take place along the border of the boundary region, and this process lay lead to the formation of larger eddies.

It is hoped that in the future it will be possible to undertake special series of observations during periods of upwelling in order to obtain better knowledge of the phenomenon and to answer many questions which now must be left open. The data from the three cruises in 1937 have, however, shown that the details of the process of upwelling are much more complicated than any previous records have revealed.

SUMMARY

Observations on three cruises off the coast of California in 1937 indicated that during the first part of a period of upwelling the light surface-water is banked up at some distance from the coast. Between the heavier upwelled water and the light offshore water a boundary region forms with a strong current parallel to the coast. This boundary region moves away from the coast, but so slowly that it acts as a barrier against the outward transport of the upwelled water. Within this a cellular circulation develops which, since the offshore boundary moves away from the coast, is being fed from below.

This appears to be the initial stage, but as the boundary region moves out conditions become more and more unstable and large eddies develop on the coastal side of the boundary. The processes are far more complicated than assumed on the basis of earlier data.

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CORRECTION

Journal of Marine Research, Vol. I, No. 1

On The Evaporation From The Oceans. By H. U. Sverdrup. Page 8.

For:

$$(11) \qquad \frac{u}{w} + 5.57 \log \frac{u}{w} = 5.5 + 5.75 \log \frac{zu}{\nu}$$

Read:

$$(11) \qquad \frac{u}{w} + \mathbf{5.75} \log \frac{u}{w} = 5.5 + 5.75 \log \frac{zu}{\nu}$$

FLUCTUATIONS IN MONTHLY SEA LEVEL ON EASTERN U. S. COAST AS RELATED TO DYNAMICS OF WESTERN NORTH ATLANTIC OCEAN

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and

Massachusetts Institute of Technology

In the determination of the pressure and velocity distributions in the oceans, very little use has heretofore been made of the observations of sea level available from the many permanent tidal stations. Insofar as the tidal stations have been interconnected by accurate leveling surveys, such observations would aid in the determination of absolute pressure and velocity distributions. Even where the tidal stations are not so connected the variations observed may still be applied in studying fluctuations of pressure and velocity. The present paper constitutes a preliminary discussion of the causes of sea level fluctuations and an exploratory investigation intended to reveal the extent to which it may be profitable to apply tidal station observations of sea level to the oceanography of the western North Atlantic.

Of the numerous permanent tidal stations available the representative stations chosen, as being apparently relatively independent of coastal influences and as being of prime importance with regard to the dynamics of the current system, were Bermuda, Charleston, Miami and Key West. See Figure 67. The stations further north are mostly located in inland waters such as Chesapeake Bay, and even the more exposed ones such as Atlantic City are flanked by a fairly broad continental shelf and are still further separated from the main current system, the Gulf Stream, by a wide band of slope water which is not free from gradient currents; hence they are omitted from this preliminary analysis. Monthly values of sea level for the four stations listed above were kindly made available by the Director of the Coast and Geodetic Survey, U. S. Department of Commerce. I have expressed these as departures from mean sea level and changed the unit to the centimeter.

Since it is desired to determine the pressure field in the water, the effect of variations in atmospheric pressure has been eliminated from the sea level departures by adding the monthly barometric departures from normal

* Contribution No. 170.

for the respective stations, expressed in centimeters of water. This correction rarely amounts to more than 4 cm. Monthly atmospheric pressures for Charleston, Miami and Key West were from "World Weather Records" and the *Monthly Weather Review;* those for Bermuda were supplied in manuscript by Lieutenant Commander H. B. F. Moorhead. The resulting reduced monthly values of sea level are given in Tables 1–4. The mean monthly values formed from these tables are plotted in Figure 68.

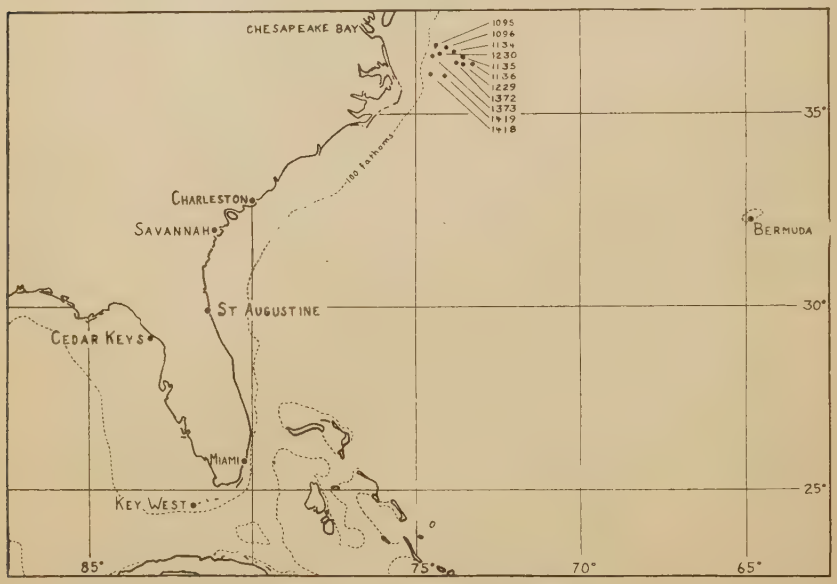

Figure 67. Locations of stations.

COMPARISON OF VARIATIONS IN MONTHLY SEA LEVEL ON EASTERN U. S. COAST AND IN OFFSHORE SEA LEVEL CALCULATED FROM OCEANOGRAPHIC DATA

It is standard oceanographic procedure to compute space variations in the geopotential (dynamic height) of the sea surface from the observed sub-surface density distribution on the assumption that at a certain deep level there are no horizontal pressure gradients. Thus one arrives at the deviations of sea level from a level surface, or surface of constant geopotential. The usual unit for expressing geopotential is the dynamic centimeter, which for practical purposes has the same *magnitude* as the linear centimeter (the error is only the percentile deviation of local gravitational acceleration from 1000). The basis for assuming no horizontal pressure gradients at a deep level is that the motion there, so far as we know, is very

slow. It is customary for deep-water calculations to assume an isobaric surface of 2000 decibars (approximately 2000 meters deep) or more as a level surface. If the deep isobaric surface is assumed to be at a *fixed* geopotential, time variations of sea level may be computed from time variations of density.

On four occasions throughout the year 1932 *Atlantis*, research vessel of the Woods Hole Oceanographic Institution, made hydrographic sections from Chesapeake Bay to Bermuda, taking temperatures and water samples to an average depth of 2100 meters. From the resulting density distributions Dietrich (*4*) has computed the elevations of the sea surface along the sections on the assumption that there are no horizontal pressure gradients

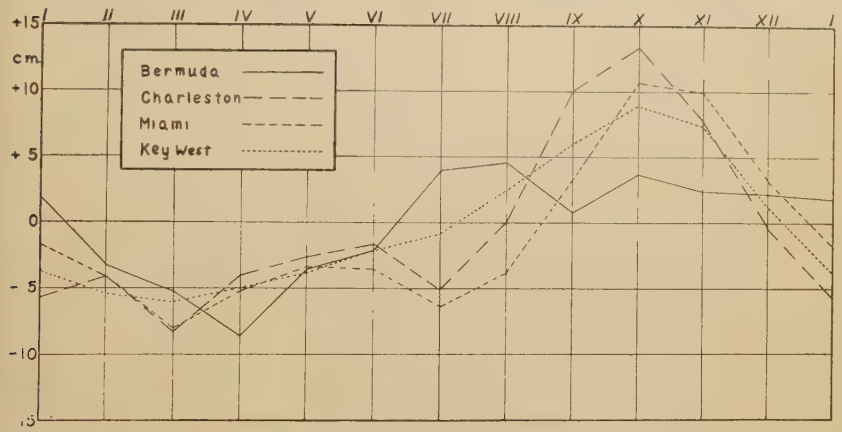

Figure 68. Annual variation of sea level reduced to normal atmospheric pressure.

in the oxygen minimum layer (250–950 meters deep). In order to obtain the height of the sea surface at the western end of the sections, where the bottom is shallower than the depth of the oxygen minimum, he has presumably assumed vanishing horizontal pressure gradients at the bottom. He has compared (abb. 15) the variation from section to section in computed height of sea level at the western end with observed monthly sea level at Atlantic City, finding rather good agreement without reducing the latter to normal atmospheric pressure.

Since the extrapolation involved in computing the elevation of the sea surface in shallow water is necessarily very unreliable, it appears to me better for the present purpose to use the height of the surface in the deep and comparatively stagnant body of water just outside the continental shelf (slope water), and to compare these computed fluctuations with monthly sea level at tidal stations relatively free from coastal influences.

Furthermore, the assumption that there are no horizontal pressure gradients in the oxygen minimum layer leads to gradient currents below this layer of the same order of magnitude as, but of opposite direction than, he finds for the surface of the Gulf Stream, a result which cannot be accepted.* Dietrich apparently now (*5*) acknowledges this criticism as just, and merely advocates the use of the oxygen minimum layer as a base level for computing the *relative* topography of isobaric surfaces. Accordingly the following recomputation is presented of the relation between fluctuations in sea level at tidal stations and those found oceanographically.

In order to eliminate as far as possible short-period oscillations, of about one day or less, as well as any observational inaccuracies, the average of two or three oceanographic stations is used to represent each occasion. From the four sections mentioned above I have used the same stations which Iselin (*8*) (p. 28) chose as representative of slope water conditions. A pair of stations made in the same region in October 1931 are also employed, making a total of five occasions. The positions of the stations are shown in Figure 67. The hydrographic data for these stations have been published in *Bulletin Hydrographique* (Appendice pour l'année 1931, and *Bulletin* pour l'année 1932).

Assuming the isobaric surface of 2200 decibars to be a level surface of fixed elevation, and assuming constant atmospheric pressure, the height of sea level has been computed from the vertical density distribution at each of these stations. The results are presented in the column designated H_a in Table 5, expressed in centimeters of dynamic height anomaly above 2200 db., which is the difference between the actual value and the dynamic height above 2200 db. for standard sea water of temperature 0°C. and salinity $35°/_{oo}$.

The mean of the five average values given in Table 5 may, for lack of further information, be regarded as representing mean sea level in the region considered. Likewise the values given in the last column of the table may be regarded as departures from mean sea level. The largest of the departures amount to 7 cm. This is about twice as great as the short-period oscillations Seiwell found at *Atlantis* station 2639 (180 miles northwest of Bermuda, maximum range during three days of 8 cm.). Hence, especially since the departures are based on two or three stations, it is unlikely that they are due to short-period oscillations. It is probable that they are largely associated with longer-period fluctuations, ones of between one week and one year.

* The possibility of a *weak* counter-current below the Gulf Stream is not doubted. While Dietrich's small slope of the sea surface across the Gulf Stream is not accepted, it may easily be that the slope resulting from using say the 2000 db. surface as a base level is too large.

Figure 69 shows the monthly values of sea level for two and a half years at Charleston and Miami, and the five oceanographic determinations. The latter show a good correlation with the tidal observations at both Charleston and Miami. The amplitude of the oceanographic variations is noticeably less than that found at Charleston, but is the same as that at Miami.

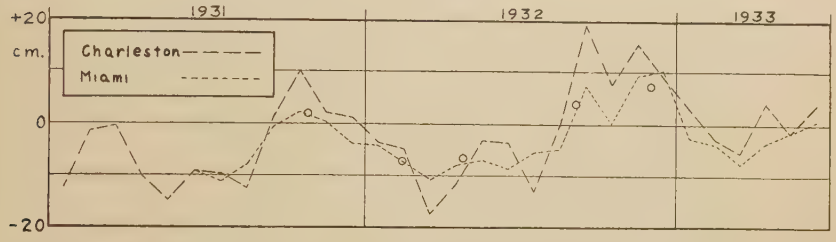

Figure 69. Monthly sea level reduced to normal atmospheric pressure at tidal stations, and five oceanographic determinations of sea level off Chesapeake Bay.

The agreement indicates that the tidal and oceanographic sea levels undergo a similar annual variation. It would not, however, be justifiable to conclude from this scanty material that deviations from the normal annual curve are similar in both sets of data. The apparent agreement shown by relatively low values in late 1931 and relatively high values in late 1932 is not altogether trustworthy, because the oceanographic value for October 1931 is based on two isolated stations instead of two or three stations chosen from a section crossing the Gulf Stream as in the other four cases, and because the western end of the section in late 1932 was somewhat further south than the other sections so that probably only station 1418 is representative of slope water conditions (cf. Iselin (*8*), p. 21).

CAUSES OF FLUCTUATIONS IN SEA LEVEL

The possible causes of long-period variations in sea level (reduced to normal pressure) at any station may be conveniently summarized as below*:

(1) *Dynamic* effects due to fluctuations in gradient currents.
(2) Effect of local winds or *Windstau.*
(3) *Local addition of mass.* This is especially effective in the tidewater of rivers, where sea level fluctuates with the river discharge. The effect of fresh water run-off is presumably negligible at the tidal stations discussed in this paper. It is also assumed that the addition

* A somewhat different classification may be found in Krümmel's "Handbuch" (*10*).

or subtraction of mass by precipitation and evaporation at the sea surface is so quickly equalized over large areas that it is of no significance locally with regard to fluctuations in monthly or weekly sea level. Change of total mass of oceans would become effective under (6).

(4) *Changes in configuration of ocean floor.* Alterations in a river channel affect sea level in its tidewater, but this is of no significance at the stations discussed in this paper. Earthquakes which alter the volume of the ocean basins would become effective under (6).

(5) *Atmospheric pressure.* The principal effect of fluctuations in atmospheric pressure is eliminated by the hydrostatic correction that has been applied to the tidal station observations. There is of course also a transient effect, but this would be of as short duration as the effect of rainfall mentioned under (3).* Application of the aforementioned correction involves the assumption that the average atmospheric pressure over the oceans is constant; this is nearly the case but introduces in the corrected values a variation amounting everywhere only to the variation in average pressure over the oceans (expressed as height of water).

(6) The *widespread* effects due to the action of any of the above at a distance or over large areas are placed together under this heading because at a given station they become indistinguishable. For lack of information this will not be discussed further.†

* The principal basis for the statements that addition of mass or increase of pressure at the surface are quickly equalized over large areas is the observed value of the "barometer factor," or average ratio of the departure from normal of sea level to the departure from normal in height of a mercurial barometer. The results for many stations give a mean value of the barometer factor which is very close to the negative of the density of mercury, indicating that there is very little lag in the adjustment of sea level to fluctuations in atmospheric pressure so as to maintain constant pressure at fixed points below the surface. For a recent brief summary concerning the barometer factor see Dietrich (*4*) (pp. 33–35).

† The action of (3), (4) and (5) in producing a widespread effect was discussed under those headings respectively. As an example of the action of (2) the following suggestion may be presented. Imagine a canal encircling the earth in the zone of westerly winds, and let there be a constriction at one part of the canal; sea level would be higher at the western end of this constriction than at the eastern end due to *Windstau.* This is essentially the case in the Southern Ocean and the constriction occurs at Drake Strait, sea level being higher on the Pacific side and lower on the Atlantic side (presumably!), accounting for the probable fact that the Pacific Ocean is generally at a higher level than the Atlantic. As a result of leveling surveys it has been found that "The average mean sea-level determination by tidal observations at Pacific Coast stations is about 0.5 meter higher than the average for the mean sea-levels at tidal stations for the Atlantic and Gulf Coasts." (*3*).

(7) *Thermal* effects.

(a) *Local thermal* effects. If a given isobaric surface below the layer affected by seasonal temperature variations, say the 200 decibar surface, maintains constant elevation, then sea level must fluctuate in accord with the heat (and, included here as a factor which affects the density of sea water, fresh water) supplied or withdrawn at the surface.

(b) *Advection.* Heat (and fresh water) may be supplied or withdrawn from a region by advection of different water masses. This refers especially to the condition maintaining coastal currents, the light water being supplied by fresh water run-off.

(8) *Astronomic tides* of long period.

It is probable that in the open ocean the long-period components of the *astronomic tide* approximate closely to the values given by the equilibrium tidal theory (*11*). The latter vanish at latitude 35° and have their maximum amplitudes at the poles as follows (the principal components only are listed and the values are for a rigid, ocean-covered earth; they are from Schureman (*16*), pp. 34, 55):

lunar fortnightly	4.1 cm.
lunisolar fortnightly	.2
lunar monthly	2.2
solar semiannual	2.0

The first three periods are of such length that they would not appear in the monthly means of sea level. Harmonic analyses for a few foreign stations give amplitudes for these three components considerably exceeding the theoretical values (Schureman (*16*), pp. 403–413), but these exceptional cases must be due to the effect of local topographical conditions on the astronomic tide. On the other hand a large semiannual component, as well as annual component which has no astronomic basis, are found at nearly all stations. At Charleston these are 9.4 and 7.4 cm. respectively. True tidal forces must therefore be of minor importance in explaining sea-level variations observed at well-exposed tidal stations.

Equilibrium tides would not produce any variations in the calculated dynamic height of the sea surface. The observed agreement between variations in dynamic height of the sea surface and in sea level at tidal stations therefore gives further evidence that the variations are not due to astronomic tides.

The *thermal* effect (produced locally or by advection) must be of importance in all temperate regions. This is substantiated by the fact that all of Marmer's (*12*) diagrams of the annual variation of sea level at tidal stations on the Atlantic and Gulf coasts of the United States (and the

Bermuda curve in Figure 68) have their maxima between June and October. At some other stations this effect is outweighed; Seattle's curve for example is below normal during April–October. The order of magnitude of the thermal effect is easily computed. An increase in temperature of the surface hundred meters from 8° to 18° at a salinity of 35‰ (roughly representing the change from February to July at the 100-meter contour off Chesapeake Bay according to Bigelow's diagrams (*2*)) would raise sea level by 19 cm. This is only slightly less than the range in sea level at Charleston shown in Figure 69, but the double maximum and the late occurrence of the chief maximum show definitely that the thermal effect is not the only one of importance.

Most writers on the subject of sea level fluctuations assume that the fluctuations are in large part due to *Windstau*, and find a close relationship with local winds. For instance Bergsten (*1*), from a comparison of monthly values of sea level at a number of tidal stations in Scandinavia and Denmark with the winds for each station, concluded definitely that all essential characteristics of the normal annual variations in sea level for those stations are caused by the local winds. Palmén, in his study of daily sea level and winds at Baltic stations (*14*), divided *Windstau* into two parts: the component of the slope of the water surface in the direction of the wind being pure *wind-effect* (*Windeffekt*), and the component at right angles being *stream-effect* (*Strömeffekt*). The latter is due to the longshore wind component, and can develop only to the extent that a longshore gradient current develops.* The gradient current is favored by a long straight coastline and by a wind that acts equally along the entire coastline. Hence this component of *Windstau* becomes to large extent indistinguishable from the dynamic effect.

For a rough indication of the annual variation of onshore wind in the vicinity of Charleston, use has been made of the monthly percentage frequency of wind direction at Savannah (*9*). Table 6 gives the average for the two periods 1891–1902 and 1923–1927 of the monthly frequencies of wind from NE, E and SE (first line), from SW, W and NW (second line),

* The distinction between wind-effect and stream-effect, which are considered here for the stationary state of motion only, may be illustrated as follows. Along a *deep* coastline a normal drift current is set up with its transport cross-wind. An onshore or offshore wind produces no *Windstau*, but a longshore wind produces *Windstau* of stream-effect type which sets up a gradient current, the amount of *Windstau* being regulated by the gradient current. Along a very *shallow* coastline frictional forces entirely outweigh the Coriolis force; the drift current is down-wind and gradient currents cannot exist. An onshore or offshore wind produces *Windstau* of the wind-effect type, a longshore wind produces no *Windstau*. In water of intermediate depth, being the normal case, both types of *Windstau* occur simultaneously (except for directly onshore, offshore, or longshore winds).

and the difference between these two (third line). This difference is intended to represent the net onshore wind. In the last line of the table are given the departures from normal of the net onshore wind. These are shown graphically in Figure 70, together with the annual variation of sea level at Charleston. The agreement between the two curves is striking. It must be remembered, however, that merely frequency of wind direction cannot give a complete description of the stress exerted by the wind on the water, especially since the latter is proportional to the second power of wind speed;* unfortunately a convenient compilation of velocities is not available. Further, both these curves bear a resemblance to the annual

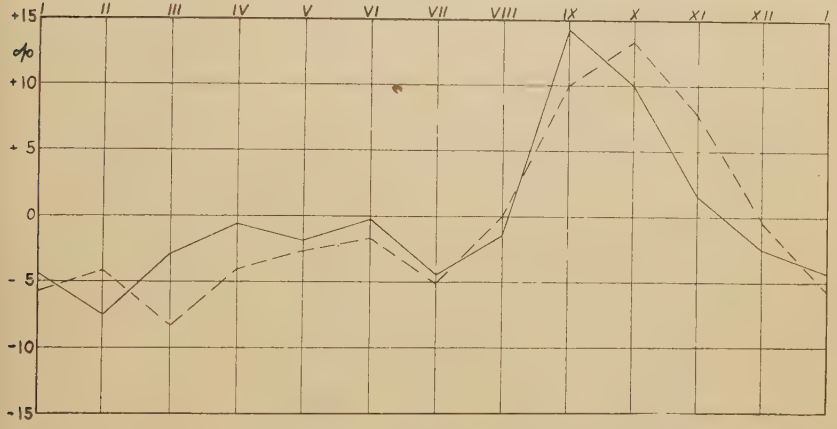

Figure 70. Full line: percentage frequency of net "onshore" winds at Savannah (from last row of Table 6). Broken line: annual variation of sea level at Charleston (reproduced from Figure 2).

variation of the northeast trades (see last section), which are ultimately largely responsible for the Gulf Stream. It is safe to conclude that the wind plays an important rôle in the annual variation of sea level at Charleston, and that part of this is the direct effect of local *Windstau*, but it could not be definitely concluded that the whole, or even major part, of the sea level variation is due to local *Windstau*.

The *dynamic* effect refers to fluctuations in sea level brought about by fluctuations in gradient currents.† A gradient current, as its name implies,

* Some authors find the power to be less than 2, as for instance Palmén (*14*), but there can be no doubt that the power is considerably greater than unity.

† An interesting example of the dynamic effect, although it does not affect mean sea level, is its well-known action on a progressive tidal wave in a canal. High water and maximum flood at any point occur simultaneously, as do low water and maxi-

requires a definite slope of the isobaric surface through each point (the slope depending only on the velocity, density and latitude at the point), and in the special case of surface gradient currents a definite slope of the sea surface is required. The dynamic effect, where the pressure distribution is produced by the motion, is to be clearly distinguished from the thermal effect, where the motion is produced by the pressure distribution. Fluctuations in sea level from the dynamic effect are to be thought of as due to fluctuations in surface gradient currents produced by forces acting on them before they enter (or after they leave) the vicinity in question.

In the region here under consideration it is to be expected that considerable dynamic effect should occur due to fluctuations in the Florida Current or Antilles Current as they leave their source.*

As was mentioned above, it is difficult to distinguish between the local stream-effect of *Windstau* and the dynamic effect resulting from its simultaneous action in the surrounding regions. In the ocean, where the action of the stream-effect is essentially wide-spread, it would seem proper to call the immediate cause of the resulting sea level fluctuations *dynamic*, although the ultimate stream-effect cause is active in the same general region.

The good agreement of the fluctuations in sea level at the oceanographic stations off Chesapeake Bay with the fluctuations at Miami—even better than with those at the nearer station Charleston—indicates that, all along the left edge of the Florida Current and Gulf Stream from the Straits of Florida to beyond Cape Hatteras, sea level varies as a unit. This strongly suggests that the dynamic effect is of prime importance in producing these fluctuations. But how much of the dynamic effect is due to variations in the current before it reaches the Florida Straits, and how much to the action of winds after leaving the Florida Straits, cannot be said. At Charleston,

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mum ebb. The dynamic effect on the tidal current produces a higher high water and a lower low water on one side of the canal than at the center, and a lower high water and higher low water on the other. For instance Marmer (*13*) has studied this difference in range for the Hudson River between Jersey City and lower Manhattan. Here the average surface current across the River at high and low water is 1.7 knots and the width 0.6 mile, giving a calculated difference in range of 0.06 foot (2 cm.), which compares favorably with the observed value of 0.08 foot.

* Since these two currents are in large part continuations of the Equatorial Current, it is of interest to mention Gallé's discussion of the rôle of fluctuations in the North Equatorial Current with respect to the annual variation of sea level observed at various tidal stations on the North Atlantic (*7*). He finds the annual variation of these two quantities closely correlated, using a time lag depending on the particular coastal region. His conclusion is that the former is not the cause of the latter, but rather that both find their direct cause in the atmospheric circulation around the North Atlantic high pressure area. This conclusion, however, cannot be regarded as final.

however, which is separated from the Gulf Stream by 50 miles of continental shelf, the monthly fluctuations are greater in magnitude (although the annual variation is of practically the same amplitude as at Miami), this increased amplitude being presumably due to local wind action.

FLUCTUATIONS IN SURFACE CURRENT DETERMINED FROM CROSS-STREAM GRADIENT OF SEA LEVEL

Regardless of the origin of the gradient currents it should be possible to compute, from sea level observed at two tidal stations separated by a body of water, the fluctuations in average surface gradient current normal to the line joining the two stations. This method was suggested by Sandström in 1903 (*15*), with especial regard to a line joining Iceland and the Norwegian coast.* But it is necessary to eliminate from the tidal station observations not only the barometric fluctuations, but also the wind-effect of *Windstau* and the effect of bottom friction on the currents near shore. These alter the sea surface near shore from the topography that would correspond to the currents, but outside the 200 meter bathymetric contour they have little effect if any.

From the agreement between the oceanographic and tidal-station determinations it may be concluded that this correction for Charleston is only a fraction of the total variation, say less than one half. As a first approximation it may be neglected altogether.

Bermuda is a station which, used in conjunction with Charleston, ought to make it possible to apply Sandström's method. An attempt was made to compare the tidal station data with the dynamic heights of the surface computed for oceanographic stations near Bermuda. Unfortunately the Bermuda tidal station was not established until November 1932, and since that time few oceanographic stations have been occupied in that vicinity. Dynamic heights computed for *Atlantis* stations within about 180 miles of Bermuda during 1931–1936 do not follow a regular annual variation nor do the ones since November 1932 agree with the monthly sea levels observed at Bermuda. This may be due to the fact that, as shown by the Chesapeake Bay to Bermuda sections figured by Iselin (*8*) (his Figs. 3–8, 10–11), the isolines of temperature and salinity have a considerable slope, usually upwards, towards Bermuda. Oft-repeated stations within a few miles of Bermuda would make it possible to gain some sort of a consistent picture of the relation between computed dynamic height of the surface and sea level observed at the tidal station.

If it is assumed, as was justified for Charleston, that sea level fluctuations observed at the Bermuda tidal station are roughly representative of the sea level fluctuations in deep water nearby, it may be that the difference of the

* So far as I know, the method has never been put into practice.

monthly departures of sea level at Bermuda and at Charleston give a rough indication of the fluctuations of the average surface gradient current between Charleston and Bermuda. The difference Bermuda—Charleston has been taken from Tables 1 and 3 for each of the 47 months available and the monthly averages (after adding 2.8 cm. to reduce the normal to zero) are plotted in Figure 71 as the full line. This indicates that the average northward surface gradient current between the two stations has its maximum in July and minimum in October, with secondary maximum in January and minimum in April. Since sea level at Bermuda is about 80 cm. higher than at Charleston (calculated from the density distribution), the

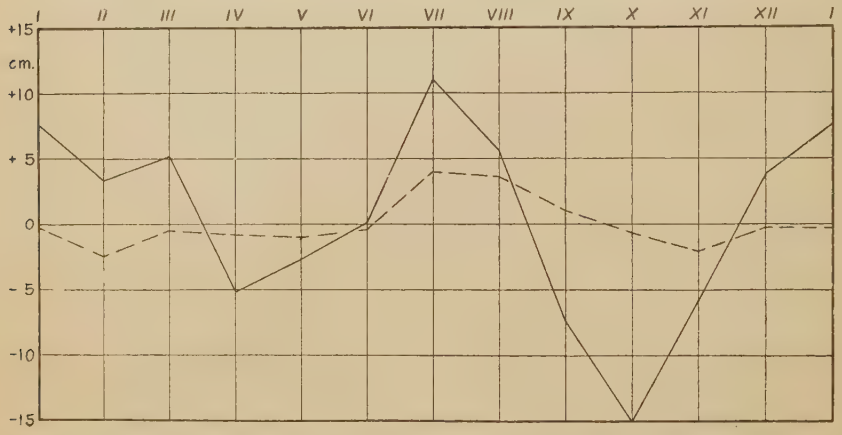

Figure 71. Annual variations of differences of sea level reduced to normal atmospheric pressure: full line for Bermuda–Charleston, broken line for KeyWest–Miami.

July and October values of +11 and −15 cm. represent deviations from normal of about 16%. Whether or not the associated fluctuations in current occur largely within the Gulf Stream cannot be said.

The results of the last paragraph are very uncertain. But it is clearly suggestive that an exact and useful application of the methods outlined may be possible if certain further observational material becomes available. At its source the Florida Current passes between Key West and Habana (Cuba), a distance of only 85 miles; further downstream it passes between Miami and the Bimini Islands of the Bahama group, a distance of only 50 miles. Since permanent tidal stations already exist at Key West and Miami, the establishment of stations at Habana and Bimini would make it possible, as Dietrich (*4*) (pp. 27–28) has already suggested, to apply this method of cross-stream gradient of sea level to two lines crossing the Florida Current.

FLUCTUATIONS IN SURFACE CURRENT DETERMINED FROM DOWNSTREAM GRADIENT OF SEA LEVEL

It has been found from leveling surveys that sea level is higher in the Gulf of Mexico than along the Atlantic coast. This head or difference in potential between the upstream and downstream ends of the Straits of Florida probably accounts for the major part of the energy of the Florida Current. On the assumption that the head accounts for all of the energy (i. e. that the Gulf of Mexico is a simple reservoir) and that the flow through the Straits is frictionless, the velocity at the downstream end of the Straits may be computed from Torricelli's theorem, $v = \sqrt{2gh}$ (v, g, h being velocity, acceleration of gravity, and head). The observed difference in level between Cedar Keys and St. Augustine is 19 cm. (*3*). These stations are rather far north from the Straits, and also their sea levels should be corrected as mentioned in the last section, but 19 cm. may be used as a rough value of the head. The resulting velocity is 193 cm. sec.$^{-1}$, which is close to the average velocity in the center of the Florida Current.

As a measure of fluctuations in head the difference of the monthly sea levels at Key West and Miami may be used. These should also be corrected, but the corrections are supposedly small due to the good exposures at these stations. The monthly differences taken from Tables 4 and 2 are shown in Table 7; this table also gives the mean monthly differences and in the last row these means reduced to a normal of zero. The last row is plotted as the broken line in Figure 71. (This curve is based on 67 monthly values, the Bermuda-Charleston curve on only 47.)

The maximum head occurs in July and the minima in November and February. Unfortunately no information concerning leveling between Key West and Miami has been published, so it is necessary to assume that the difference between these two stations is the same as between Cedar Keys and St. Augustine, namely 19 cm. On this basis the July and November values represent deviations from normal head of 16%. In terms of velocity this becomes 8%.

The Florida Current supplies part of the Gulf Stream, which in turn accounts for most of the surface gradient current between Charleston and Bermuda, so it is to be expected that the fluctuations in downstream gradient of sea level from Key West to Miami are related to the fluctuations in cross-stream gradient from Bermuda to Charleston. The agreement found in magnitude (8% and 16% in terms of velocity) and in character of the annual variation (Figure 71) is far from close, yet it is sufficient* to indicate

* The correlation coefficient between 47 monthly values of Bermuda—Charleston and Key West—Miami is $+ .32 \pm .09$ (second term is probable error). But Charleston and Miami are quite similar, and the correlation between Bermuda—Miami and Key West—Miami would be $+ .50$ if the three stations were entirely independent and

the potential value of the two methods used and to lend weight to Table 7 as giving roughly the monthly deviations from normal in strength of the Florida Current. The mean monthly values shown in Figure 71 also resemble the strength of the northeast trades in the same or preceding month; some such relationship is to be expected since the trades provide the supply of water to the Gulf of Mexico as well as to the Antilles Current. For the trades in the region 15°–25° N. and 25°–40° W. Gallé (*6*) gives:

month	*I*	*II*	*III*	*IV*	*V*	*VI*	*VII*	*VIII*	*IX*	*X*	*XI*	*XII*
wind in m.p.s.	5.5	5.2	4.6	4.7	5.8	6.2	6.1	5.3	5.3	4.5	4.6	4.7

In conclusion it is to be recommended that permanent tidal stations be established at Habana and Bimini,* and that Miami and Key West be included in the network of precise leveling. With the further use of suitably planned oceanographic observations and of the orginal records from coastal meteorological stations it should be possible to gain a fairly complete knowledge of the causes of the sea level fluctuations in the region discussed in this paper. This knowledge could be used to determine accurately, from tidal and meteorological observations alone, the fluctuations in the surface velocity of the Florida Current.

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the fluctuations had the same standard deviations. Actually the latter correlation has the value + .49 ± .07. Hence at first sight it does not appear that cross-stream and downstream components of sea level lead independently to similar fluctuations of the current. However, the following analysis indicates that the coefficient + .49 is significant.

In the case of three series of numbers a, b, c the correlation between a–c and b–c depends merely on the standard deviations of a, b, c and the three individual correlations between them. Let a represent Key West, b Bermuda, and c Miami. In the present case the standard deviations are *not* the same, but are

$$\sigma_a = 7.08 \text{ cm.}, \quad \sigma_b = 7.26, \quad \sigma_c = 7.74,$$

and the correlation between Key West and Miami is *high*, namely $r_{ac} = +.887$. If there were no significance in the correlation between a–c and b–c it would have the value given by assuming b entirely independent of a and c, namely

$$r = \frac{\sigma_c^2 - \sigma_a \sigma_c r_{ac}}{\sqrt{(\sigma_a^2 - 2\sigma_a \sigma_c r_{ac} + \sigma_c^2)\,(\sigma_b^2 + \sigma_c^2)}} = +.30.$$

But the actual coefficient +.49 is significantly larger.

* The Woods Hole Oceanographic Institution, with the cooperation of the Coast and Geodetic Survey, has just established a permanent tidal station on Cat Cay, one of the Bimini Islands, which was set in operation on March 1, 1938.

SUMMARY

1. Variations in the dynamic height of the sea surface observed on five occasions during 1931–1932 in the slope water on the left side of the Gulf Stream off Chesapeake Bay are found to agree closely with monthly sea level reduced to normal atmospheric pressure observed at Charleston and Miami.

2. The possible causes of these long-period fluctuations are discussed and it is concluded that only the following can be of importance:

(1) *Dynamic* effect due to variations in gradient currents, in particular the Gulf Stream
  (a) caused by variations in the Florida Current and Antilles Current at their source.
  (b) caused by the action of wind (stream-effect).
(2) *Windstau* (wind-effect).
(6) *Widespread* effects.
(7) *Thermal* effects.

It may be concluded tentatively that (1) and (7) are of most importance for Miami and for the deep-water variations, while (2) also exerts a large influence at Charleston.

3. Means are discussed for determining fluctuations in the average surface velocity, especially of the Florida Current, merely from observations at tidal stations of the cross-stream gradient of sea level and from meteorological observations.

4. Similar means are discussed using the downstream gradient of sea level. A preliminary application of this method results in the monthly values in Table 7, which give a rough indication of fluctuations in the Florida Current.

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TABLE 1

DEPARTURES FROM NORMAL IN CM. OF MONTHLY SEA LEVEL AT ST. GEORGES, BERMUDA, REDUCED TO SEA LEVEL PRESSURE OF 1018.0 MB.

(Normal sea level is the average for 1933–1935, being staff reading of 4.19 feet; zero on staff is 8.95 feet below Bench Mark 1.)

Month	*I*	*II*	*III*	*IV*	*V*	*VI*	*VII*	*VIII*	*IX*	*X*	*XI*	*XII*
1932	—	—	—	—	—	—	—	—	—	—	− 1.0	− 1.7
1933			− 3.1	−13.6	− 3.6	+ 9.3	0.0	−10.2	− 6.4	+ 9.2	+ 2.2	+ 0.6
1934	− 0.4	− 3.7	− 8.3	− 6.7	+ 3.1	− 8.4	+ 6.0	+21.4	+13.3	− 5.5	− 0.9	+ 7.8
1935	+ 9.3	+ 2.3	+ 0.9	− 6.8	− 1.8	− 0.6	+ 2.9	+ 4.9	− 3.6	+ 0.2	+ 0.3	+ 4.4
1936	− 3.6	− 8.6	−10.1	− 7.5	−11.7	− 8.8	+ 6.9	+ 2.4	0.0	+10.8	+11.6	—
Means	+ 1.8	− 3.3	− 5.2	− 8.6	− 3.5	− 2.1	+ 4.0	+ 4.6	+ 0.8	+ 3.7	+ 2.4	+ 2.2

TABLE 2

DEPARTURES FROM NORMAL IN CM. OF MONTHLY SEA LEVEL AT MIAMI BEACH REDUCED TO SEA LEVEL PRESSURE OF 30.04 INCHES

(Normal sea level is the average for 1932–1936, being staff reading of 3.42 feet; zero on staff is 9.48 feet below Bench Mark 6.)

Month	*I*	*II*	*III*	*IV*	*V*	*VI*	*VII*	*VIII*	*IX*	*X*	*XI*	*XII*	*Year*
1931	—	—	—	—	—	− 8.9	−11.1	− 7.8	− 0.5	+ 2.3	+ 0.4	− 3.7	—
1932	− 4.1	− 7.4	−10.8	− 8.0	− 6.9	− 8.5	− 5.5	− 4.8	+ 7.6	+ 0.3	+ 9.5	+10.5	− 2.3
1933	− 2.5	− 3.6	− 7.7	− 3.6	− 1.4	+ 0.7	− 4.6	− 0.6	+ 3.0	+19.3	+15.9	− 2.6	+ 1.0
1934	− 2.7	− 7.6	− 8.9	− 5.8	− 8.3	− 4.8	− 6.7	− 3.9	+ 3.5	+13.2	+ 7.4	+ 1.3	− 1.9
1935	+ 4.2	− 0.6	− 2.9	− 0.8	− 6.8	− 4.1	− 5.8	− 4.8	− 3.2	+16.1	+18.0	+ 4.2	+ 1.1
1936	− 3.2	− 1.5	− 9.9	− 7.7	+ 6.7	+ 4.4	− 4.6	− 0.8	+ 9.5	+12.8	+ 8.9	+ 9.3	+ 2.0
Means	− 1.7	− 4.1	− 8.0	− 5.2	− 3.3	− 3.5	− 6.4	− 3.8	+ 3.3	+10.7	+10.0	+ 3.2	—

TABLE 3

DEPARTURES FROM NORMAL IN CM. OF MONTHLY SEA LEVEL AT CHARLESTON REDUCED TO SEA LEVEL PRESSURE OF 30.07 INCHES

(Normal sea level is the average for 1922–1936, being staff reading of 4.96 feet; zero on staff is 14.22 feet below Bench Mark 1.)

Month	*I*	*II*	*III*	*IV*	*V*	*VI*	*VII*	*VIII*	*IX*	*X*	*XI*	*XII*	*Year*
1921	—	—	—	—	—	—	—	—	—	+13.6	+ 7.2	+ 3.0	—
1922	− 7.1	+ 0.9	− 1.4	− 2.7	+ 0.7	− 0.1	− 3.6	− 1.9	+18.1	+10.0	+ 7.9	− 6.2	+ 1.2
1923	− 4.4	− 0.7	−10.4	− 2.7	+ 2.9	− 3.7	− 7.1	−10.6	+ 3.5	+14.4	+ 8.7	− 6.5	− 1.4
1924	− 9.9	− 2.6	−11.5	− 6.8	−10.9	− 7.3	− 0.7	+ 3.2	+18.0	+23.3	+ 9.6	− 0.5	+ 0.3
1925	+ 6.6	− 0.2	− 1.3	+ 0.8	+ 0.7	+ 1.8	− 0.2	+12.9	+10.3	+ 0.2	+10.3	−12.3	+ 2.5
1926	−16.9	−21.7	−16.1	− 9.8	− 5.8	− 2.5	− 4.7	+ 2.5	+17.1	+ 6.9	− 2.1	− 3.8	− 4.7
1927	− 8.7	−11.8	− 0.2	+ 0.7	− 6.0	+ 1.4	− 6.1	− 1.1	+ 3.0	+ 4.5	+ 4.5	+ 1.6	− 1.5
1928	−19.5	−10.2	− 9.6	− 8.9	+ 0.5	− 8.4	− 4.0	+ 3.7	+ 8.4	+17.8	+ 8.7	+ 8.7	− 1.1
1929	−13.3	+ 1.9	−10.7	− 4.9	− 3.9	+ 8.4	+ 0.5	− 1.1	+11.9	+ 6.2	+ 7.4	− 7.2	− 0.4
1930	+ 9.9	+ 3.0	−15.1	− 0.9	− 7.8	− 3.7	−10.5	− 0.6	− 3.7	+13.8	+ 4.3	− 4.1	− 1.3
1931	−12.6	− 1.5	− 0.5	−10.4	−14.7	− 9.0	− 9.8	−12.5	+ 1.5	+10.2	+ 2.2	+ 1.3	− 4.6
1932	− 3.4	− 4.6	−17.5	−11.5	− 3.1	− 3.5	−12.9	+ 0.3	+19.4	+ 7.7	+15.8	+ 9.4	− 0.3
1933	+ 3.2	− 2.8	− 5.5	+ 4.2	− 1.5	+ 3.7	+ 2.7	+ 9.1	+ 9.0	+23.2	+ 7.1	− 2.1	+ 4.2
1934	− 6.6	− 5.4	− 7.3	− 5.4	− 1.0	− 9.2	− 8.5	− 7.3	+ 7.8	+21.9	+ 6.5	+ 1.0	− 1.1
1935	+ 8.3	− 4.1	− 3.1	− 0.9	+ 1.3	− 1.0	− 5.0	+ 1.0	+10.7	+19.6	+21.7	− 1.3	+ 3.9
1936	−10.7	− 2.0	−14.2	− 0.5	+ 9.1	+ 8.7	− 6.1	+ 4.5	+16.7	+21.6	+ 4.2	+12.6	+ 3.7
Means	− 5.7	− 4.1	− 8.3	− 4.0	− 2.6	− 1.6	− 5.1	+ 0.1	+10.1	+13.4	+ 7.8	− 0.4	—

TABLE 4

DEPARTURE FROM NORMAL IN CM. OF MONTHLY SEA LEVEL AT KEY WEST REDUCED TO SEA LEVEL PRESSURE OF 30.02 INCHES

(Normal sea level is the average for 1913–1936, being staff reading of 4.93 feet; zero on staff is 13.06 feet below Bench Mark 29.)

Month	*I*	*II*	*III*	*IV*	*V*	*VI*	*VII*	*VIII*	*IX*	*X*	*XI*	*XII*	*Year*
1913	− 4.4	− 1.9	− 5.2	− 0.8	− 1.3	+ 2.0	+ 0.4	+ 6.9	+ 6.0	+ 7.3	+ 3.6	− 1.2	+ 1.0
1914	− 5.3	− 9.2	−11.4	− 7.9	+ 6.0	+ 1.2	− 4.3	+ 8.3	+10.3	+ 1.5	+ 2.3	+ 2.2	− 0.5
1915	− 1.6	− 2.2	− 4.6	− 5.2	− 9.9	− 5.7	− 3.3	− 1.1	+ 7.9	+ 8.5	+ 5.7	− 3.4	− 1.2
1916	− 3.0	− 2.5	− 9.9	− 7.2	− 1.8	− 7.6	− 0.7	+ 2.1	+ 6.2	+13.4	+10.9	+ 0.2	0.0
1917	− 7.6	− 4.3	+ 8.2	+ 2.4	− 1.3	− 2.4	− 4.6	− 5.3	+ 8.7	+16.3	+18.6	+ 7.9	+ 3.0
1918	− 5.4	− 9.6	− 5.4	− 0.4	+ 0.6	− 4.4	− 1.7	+ 2.0	+ 3.6	+ 1.5	+ 5.4	+ 4.3	− 0.8
1919	− 6.1	−11.4	− 4.1	− 7.7	− 8.4	− 2.0	+ 0.7	+ 2.3	+ 6.3	+12.4	+ 8.2	+ 1.3	− 0.7
1920	− 8.3	− 6.3	−18.5	−12.2	− 8.2	− 1.0	+ 0.7	0.0	+ 4.4	+ 7.3	+ 4.9	− 0.6	− 3.2
1921	− 1.6	− 5.5	− 3.2	0.0	− 1.4	+ 4.1	+ 0.4	+ 1.9	+10.5	+11.4	+ 3.2	+ 0.3	+ 1.7
1922	− 6.1	− 0.4	− 1.4	− 3.0	− 3.9	− 0.1	+ 4.2	+ 2.7	+ 8.4	+ 6.2	+ 3.4	− 4.6	+ 0.4
1923	− 3.9	− 4.7	− 2.3	− 2.3	− 0.2	− 1.7	− 4.4	− 4.0	+ 2.2	+ 4.1	+ 2.0	− 6.8	− 1.8
1924	− 5.4	− 3.8	− 7.8	−14.2	−10.9	− 4.6	+ 1.7	+ 4.6	+ 6.3	+ 8.2	+ 7.9	− 9.5	− 2.3
1925	− 7.7	− 4.0	− 6.4	−14.5	− 4.3	− 3.9	− 2.6	+ 3.7	+ 5.8	+ 5.1	+10.4	+ 1.1	− 1.4
1926	− 9.6	−16.0	−14.7	− 8.2	− 8.1	− 5.7	− 5.2	+ 5.8	+ 7.8	+10.3	+ 4.6	− 2.1	− 3.4
1927	+ 0.1	− 1.6	− 1.2	− 2.7	− 3.7	− 1.3	− 3.6	+ 2.8	+ 0.7	+ 6.8	+12.1	+ 4.1	+ 1.0
1928	− 4.1	−15.4	−10.4	−11.6	− 5.7	− 2.0	+ 0.9	+ 4.2	+ 5.4	+ 9.7	+ 9.6	+ 5.5	− 1.2
1929	− 7.6	− 1.8	− 6.5	− 0.9	− 1.4	− 2.1	+ 3.0	+ 3.1	+ 3.6	+ 3.7	+ 1.6	+ 2.5	− 0.2
1930	+ 3.8	+ 1.6	− 4.9	− 2.4	− 1.4	− 3.9	− 0.2	+ 6.0	+ 2.6	+ 8.6	+ 4.3	− 3.5	+ 0.9
1931	−10.7	− 9.1	− 4.4	− 4.4	−14.6	− 7.4	− 5.7	− 3.4	+ 3.7	+ 3.8	+ 5.6	+ 0.8	− 3.8
1932	+ 0.4	− 6.3	− 9.3	− 6.3	− 3.8	− 4.2	− 0.6	+ 2.1	+ 6.3	+ 5.1	+ 9.8	+11.6	+ 0.4
1933	+ 0.2	− 2.1	− 7.5	− 1.8	+ 1.3	+ 4.3	+ 2.4	+ 6.6	+ 9.0	+17.7	+13.8	+ 0.4	+ 3.7
1934	− 0.9	− 8.5	− 6.1	− 4.7	− 8.6	− 4.2	− 1.8	0.0	+ 9.2	+12.8	+ 8.2	+ 5.8	+ 0.1
1935	+ 4.5	− 3.3	− 0.2	− 0.3	− 4.1	− 2.6	+ 1.5	+ 4.7	+ 1.1	+17.5	+12.5	+ 2.4	+ 2.8
1936	− 2.0	− 1.2	− 7.8	− 4.8	+ 5.4	+ 4.7	+ 4.3	+ 3.4	+11.3	+17.1	+12.0	+10.8	+ 4.4
Means	− 3.8	− 5.4	− 6.0	− 5.0	− 3.7	− 2.1	− 0.8	+ 2.5	+ 6.1	+ 9.0	+ 7.5	+ 1.2	—

TABLE 5

DYNAMIC HEIGHT OF SURFACE AT *Atlantis* SLOPE WATER STATIONS

Station	*Date*	*L* *North*	λ *West*	H_a *dyn. cm.*	*Average*	*Departure from 131*
1095	X, 25, 1931	36°41′	74°21′	136		
1096		36°38′	74°03′	130	133	+2
1134	II, 12, 1932	36°31′	73°50′	120		
1135		36°24′	73°29′	125		
1136		36°13′	73°16′	128	124	−7
1229	IV, 22	36°12′	73°32′	125		
1230		36°27′	74°15′	124	124½	−6½
1372	IX, 2	36°16′	73°46′	137		
1373	3	36°25′	74°28′	133	135	+4
1418	XI, 30	35°57′	74°32′	133		
1419	XII, 1	35°55′	74°07′	144	138½	+7½
				mean	131	

TABLE 6

MONTHLY PERCENTAGE FREQUENCIES OF WINDS AT SAVANNAH

	I	*II*	*III*	*IV*	*V*	*VI*	*VII*	*VIII*	*IX*	*X*	*XI*	*XII*	*Mean*
NE–SE	9.5	8.3	9.7	10.0	9.7	10.3	8.7	10.7	18.5	16.3	12.0	10.0	—
SW–NW	16.5	18.5	15.3	13.3	14.2	13.2	15.8	14.8	6.8	8.8	13.2	15.2	—
Difference (net "on-shore" component)	−7.0	−10.2	−5.6	−3.3	−4.5	−2.9	−7.1	−4.1	+11.7	+ 7.5	−1.2	−5.2	−2.7
Departure from −2.7	−4.3	− 7.5	−2.9	−0.6	−1.8	−0.2	−4.4	−1.4	+14.4	+10.2	+1.5	−2.5	—

TABLE 7

MONTHLY SEA LEVEL DIFFERENCE IN CM., KEY WEST–MIAMI

(Values in Table 4 minus values in Table 2)

Month	*I*	*II*	*III*	*IV*	*V*	*VI*	*VII*	*VIII*	*IX*	*X*	*XI*	*XII*	*Year*
1931	—	—	—	—	—	+ 1.5	+ 5.4	+ 4.4	+ 4.2	+ 1.5	+ 5.2	+ 4.5	—
1932	+ 4.5	+ 1.1	+ 1.5	+ 1.7	+ 3.1	+ 4.3	+ 4.9	+ 6.9	− 1.3	+ 4.8	+ 0.3	+ 1.1	+ 2.6
1933	+ 2.7	+ 1.5	+ 0.2	+ 1.8	+ 2.7	+ 3.6	+ 7.0	+ 7.2	+ 6.0	− 1.6	− 2.1	+ 3.0	+ 2.6
1934	+ 1.8	− 0.9	+ 2.8	+ 1.1	− 0.3	+ 0.6	+ 4.9	+ 3.9	+ 5.7	− 0.4	+ 0.8	+ 4.5	+ 2.0
1935	+ 0.3	− 2.7	+ 2.7	+ 0.5	+ 2.7	+ 1.5	+ 7.3	+ 9.5	+ 4.3	+ 1.4	− 5.5	− 1.8	+ 1.7
1936	+ 1.2	+ 0.3	+ 2.1	+ 2.9	− 1.3	+ 0.3	+ 8.9	+ 4.2	+ 1.8	+ 4.3	+ 3.1	+ 1.5	+ 2.4
Means	+ 2.1	− 0.1	+ 1.9	+ 1.6	+ 1.4	+ 2.0	+ 6.4	+ 6.0	+ 3.4	+ 1.7	+ 0.3	+ 2.1	+ 2.4
−2.4	− 0.3	− 2.5	− 0.5	− 0.8	− 1.0	− 0.4	+ 4.0	+ 3.6	+ 1.0	− 0.7	− 2.1	− 0.3	—

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